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Structure and Depositional Environment of the Vendian Complex in the Southeastern White Sea Area

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Abstract—The subdivision and correlation scheme of the Vendian Complex is revised based on new data from outcrops along the eastern margin of the Baltic Shield (the southeastern White Sea area) and recently drilled boreholes. According to the established cyclic architecture of the sequence, the Vendian Complex is divided into the Lyamtsa, Verkhovka, Zimnie Gory, and Erga formations. Principal lithologic units of these subdivisions are described, and their facies relations and depositional environment are considered. It is suggested that the Vendian Complex accumulated in offshore settings as a result of southwestern progradation of a deltaic depositional system.

Key words: Vendian, stratigraphy, depositional environment, southeastern White Sea area.

INTRODUCTION

The Ediacara-type fossil biota of the southeastern White Sea area is the most informative one because of its high taxonomic diversity and extraordinary preservation (Fedonkin, 1981, 1990; Ivantsov, 1999, 2001; Grazhdankin, 2000). The Vendian Complex hosting these fossils is composed of sandy to clayey deposits exposed in relatively small and isolated outcrops each characterizing a part of the whole succession that is variably subdivided by different researchers. Moreover, the deposits have never been described in necessary details, which would properly characterize their depositional environment and settings and enable a correct understanding of stratigraphic and facies-controlled distribution of fossils in the sections. As a result, the insufficient data on habitat and burial conditions of organisms represent a serious impediment, when one is going to seek for new localities of organic remains.

The Vendian Complex of the southeastern White Sea area is up to 550 m thick and rests almost horizontally (the average dip angle not more than 0.2°) on the platform crystalline basement and Upper Riphean sediments filling the grabens and deep troughs in the latter (Stankovsky *et al.*, 1972, 1983; Yakobson *et al.*, 1991). Along the Baltic Shield eastern margin, Vendian deposits are exposed in river valleys crosscutting the Onega Ridge (Lyamtsa and Purnema rivers), a western part of the Dvina glint (Nizhma, Agma, Syuzma, Verkhovka, Solza, and Kinzhuga rivers), and the western White Sea-Kuloi glint (Torozhma and Zolotitsa rivers). In addition, they crop out along the Onega, Letnii, and Zimnii coasts of the White Sea. Toward the northeast, east, and southeast, Vendian strata plunge deeply under Paleozoic deposits of the Mezen syneclise.

The main objective of this work is to describe lithology of the Vendian Complex of the southeastern White Sea area, to revise the local scheme of its subdivision and correlation, and to interpret the depositional environment of corresponding sediments. The study is based on the bed-by-bed description of more than 150 outcrops in the Onega Peninsula and Kuloi Plateau and on investigation results of core samples from four boreholes drilled in the Onega Peninsula by the Syuzma Party of AO "Arkhgeoldobycha" in 1993 to 1996. In addition, the reference borehole sections described by E.A. Kalberg, A.I. Lebedintsev, and A.F. Stankovsky have been reexamined.

INVESTIGATION HISTORY

Deposits of the southeastern White Sea area, which are attributed now to the Vendian Complex, were comprehensively studied before the World War II by Kalberg who established in 1936–1937 that they bear organic remains originally identified as plant impressions and worm trails (the report of 1940 by Kalberg and Ershova). At that time, the clayey-sandy sequence bearing these remains and exposed in river valleys and in coastal cliffs of the White Sea was attributed to the Upper Devonian (Zekkel', 1939; Kalberg, 1940). In the period of 1947 to 1949, the Northern Geological Service drilled the boreholes Nenoksa, Arkhangelsk, and Ust-Pinega, which penetrated through the sedimentary cover of the region down to the crystalline basement. In 1950, Kalberg compared her field observations with data on these boreholes and subdivided the sequence into the Lyamtsa, Arkhangelsk, Verkhovka, Zimnie Gory, and Yagry formations (Fig. 1) based on predomi-

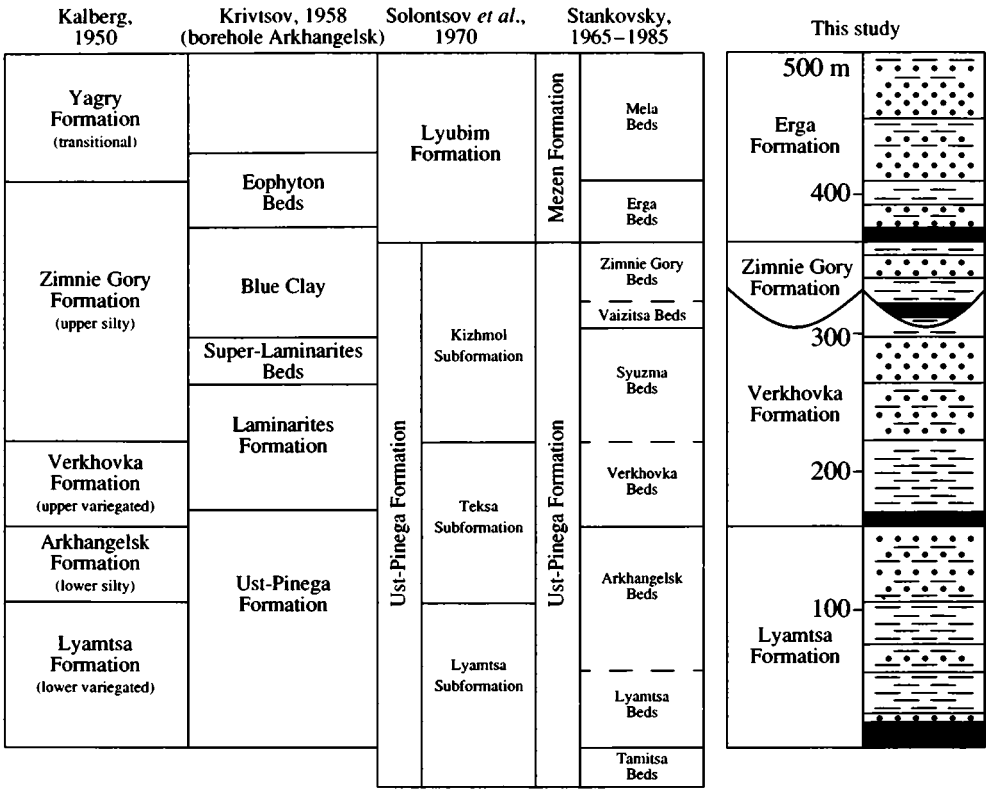


Fig. 1. Comparison of stratigraphic schemes suggested for Vendian deposits in the southeastern White Sea area (symbols as in Fig. 2).

nant coloration and petrographic composition of their rocks.

After the World War II, data of deep drilling in the East European platform showed that age of the oldest deposits in sedimentary cover must be revised, and in the northwestern and western platform areas there were distinguished the Vendian Complex of the terminal Precambrian and the overlying Baltic Group of the Lower Cambrian (Sokolov, 1952). In the 1950s, this stimulated reexamination of oldest deposits in the southeastern White Sea area, where they were found to bear remains of organic-walled tubular fossils resembling *Sabellidites cambriensis* Yanichevsky from Lower Cambrian deposits (Igolkina, 1956, 1959; Zoricheva, 1963). As a result, sedimentary strata with these remains were included into the Baltic Group, while the other ones the lacking fossils were attributed by Krivtsov (1958) to the Ust-Pinega Formation of the Proterozoic, the stratotype of which was distinguished in the depth interval of 535.75–353.60 m of the borehole Arkhangelsk (Fig. 1). Later on, it was established that *Sabellidites*-like fossils are characteristic of the Neoproterozoic deposits in general and are not suitable for stratigraphic subdivision and correlation (Stankovsky *et al.*, 1990; Gnilovskaya, 1996; Gnilovskaya *et al.*, 2000).

In 1962, the Interdepartmental Conference on Elaboration of Unified Stratigraphic Schemes for the Upper Precambrian and Paleozoic of the Russian Platform conclusively established the Vendian Complex as a specific post-Riphean stratigraphic subdivision. By that time, Igolkina (1959) described a rhythmic succession of beds exposed in the southeastern White Sea area that enabled its correlation with the Vendian stratotype and Redkino Formation of the Moscow syneclise (Aksenov and Igolkina, 1969; Aksenov and Volkova, 1969). The correlation was based on cyclic stratigraphy with reference to marker horizons of volcanic ash and distribution of organic-walled microfossils. Solontsov and Aksenov suggested a correlation of the Vendian Complex of the southeastern White Sea area with the Ust-Pinega and Lyubim formations of central and northern platform areas (Solontsov and Aksenov, 1970; Solontsov *et al.*, 1970; Aksenov *et al.*, 1978). The Ust-Pinega Formation was regarded therewith in the range established by A.I. Krivtsov in the eponymous borehole section and corresponding to the depth interval of 804–520 m (Fig. 1).

The results of geological survey (1965–1985) carried out in the southeastern White Sea area under guidance of Stankovsky showed that the Vendian succession differs here in lithologic and facies characteristics from that described in the central platform areas. In particular, the Ust-Pinega Formation as distinguished by

Solontsov and Aksenov is overlain here by a succession of interbedded reddish brown silty sandstones and maroon brown clays differing in coloration from green rocks of the Lyubim Formation. In addition, perfect exposures and numerous borehole sections exhibit coarsening-upward regressive cycles (Stankovsky *et al.*, 1981, 1985) in contrast to fining-upward transgressive cycles of the Ust-Pinega sedimentary sequence as argued by Solontsov and Aksenov. According to the results obtained, the Ust-Pinega Formation has been divided into members, some of which retained names originally given by Kalberg (Fig. 1).

In 1983, the session of ISC Bureau held in Syktyvkar adopted a stratigraphic scheme for Precambrian deposits of the southeastern White Sea area, in which the Vendian Complex is divided into the Ust-Pinega, Mezen, and Padun formations (Dedeev and Keller, 1986). That scheme ignores, however, the results obtained by Stankovsky and having been of a high practical and scientific importance until present. The Ust-Pinega Formation has been adopted as revised by Solontsov and Aksenov, whereas the Mezen and Padun formations have been proposed without reference to stratotypes. The age of the Padun Formation is problematic.

LITHOLOGIC TYPES OF DEPOSITS

The Vendian succession is composed of clayey, silty, and sandy deposits. Despite the observable lithologic variability and diverse character of bedding and lamination, the deposits are divisible into a limited number of lithologic types. Each of the latter characterizes natural sedimentary bodies (beds) displaying a persistent combination of features, which are indicative of time, genesis, and depositional settings of a given sediment. In addition to the lithological types described below, there are conglomerate, gravelstone, and volcaniclastic beds in the succession, and the carbonate, sulfate, and ironstone concretions.

Distinguished among the Vendian rocks are the following eight lithologic types:

(1) *Thin-bedded mudstone* constituting beds and members (up to 20 m thick); the lamination is visualized by silty laminae, as thin sometimes as one grain, by fine brown sapropel-like films, and by thin interbeds of volcanic ash (Plate I, fig. 2). The latter are variably thick, from 1 or 2 mm to 10 cm. Particular 1-m-thick intervals of mudstone sequence may enclose 4 to 6 ash beds.

(2) *Interbedded siltstone and mudstone* may represent intervals from few centimeters to 4 or 5 m thick; laminae of both rock types are 2 to 3 mm thick in average (Plate I, fig. 3). Siltstone laminae with sharp lower boundaries show fine graded bedding and occasionally fine cross-lamination (Plate I, fig. 5). In general, intercalated siltstone and mudstone laminae are of equal thickness, although one of the laminae type (e.g., fine-bedded siltstone, Plate I, fig. 5) can be thicker than the other one. Sandstone interbeds (5 to 10 mm thick) locally appearing as a third component are considerably thicker than mudstone laminae (1 to 2 mm); they exhibit wavy top surface and are discontinous.

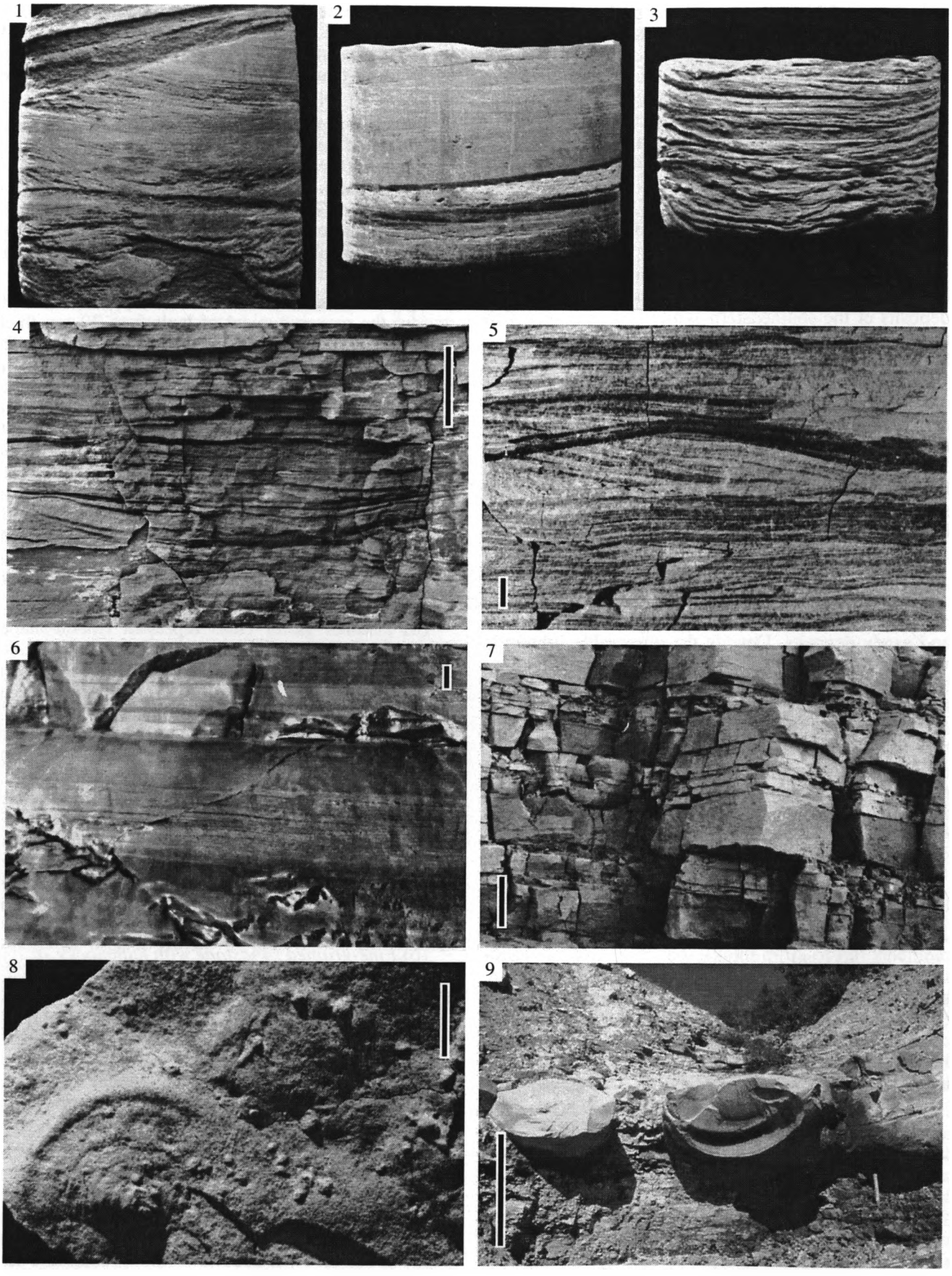
(3) *Fine-grained sandstone with thin horizontal bedding*; this is a type rock of interbeds (5 to 20 cm) or beds (up to 0.5 m thick) within the sequence of interbedding siltstone and mudstone (Plate I, fig. 7). Sole planes of beds are decorated with tool or drag marks and with flute casts, all having a uniform trend. Also typical of sole planes are load casts and accumulations of small mudstone pebbles.

(4) *Fine-grained sandstone with ripple cross-lamination*; the rock is typical of beds 5 to 20 cm thick, which are widespread in the Vendian sequence. The beds are flaggy because of multidirectional cross lamination, and their top surface is wavy (Plate I, fig. 1). They split into flags along shale partings less than 1 mm thick (films on bedding planes), which may grade into thicker shale laminae (some over 10 mm thick). Sole planes exhibit multidirectional tool marks, flute casts, and swing marks. The latter correspond to arcuate scratches and grooves produced by the bottom-anchored objects (e.g., by algae) when they swayed from side to side in turbulent flow (Plate I, fig. 8). The swing marks are either bow-shaped or bimodal in configuration.

(5) *Isolated gutter casts* represent ribbon-shaped thin-bedded sand bodies with erosional base, which originated when sandy material infilled the wide gutters in muddy substratum (Plate II, fig. 1). The casts are up to 0.3–0.4 m thick and up to 1 m wide. Their length is unknown, though steep pinch-outs are frequently visible. The cast formation was associated with scouring and liquefaction of mudstone-siltstone sequence that encloses these bodies (arguments in favor are the erosional contacts and soft deformation of the host deposits). Consequently, the gutters originated under influence of sand-saturated flows, which eroded the non-compacted muddy substratum and became immediately cast by coarser sediments. This is evidenced, in particular, by steep and "overhanging" gutter walls.

(6) *Fine-grained sandstone with hummocky stratification*; these interbeds in mudstone-siltstone sequence are 7 to 10 cm thick and have smooth erosional soles (Plate II, fig. 2). The interbeds consist of several undulatory "series," each forming the low-amplitude (up to 6 cm) hummocks, the wavelength of which can be as great as 1 m. The hummocks are erosional in origin, because stratification of underlying series is truncated under variable angles.

(7) *Fine- to medium-grained sandstone with planar lamination* constitutes beds (0.1 to 0.5 m thick) and members (up to 10 m thick) extending for a distance of several kilometers (Plate II, fig. 3). Because of planar to gently undulating lamination, the rocks are of a regular flaggy structure. Their beds have smooth soles and bed-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ding planes marked by current-formed parting lineation.

(8) *Medium- to coarse-grained sandstone with multistoried cross-lamination* is characteristic of lenticular beds 2 to 4 m thick, which have erosional soles and represent casts of overflow and distributary channels. The overflow channel casts are up to 10 m wide, having basal packages (up to 30 cm thick) of planar-laminated sandstone with diverse erosion marks at the sole (Plate II, fig. 5), followed by sandstone packages (10 to 15 cm thick) with unidirectional multistoried cross-lamination. Separate cross-laminated series are few centimeters thick in this case. The distributary-channel casts are wider, although their exact width is undeterminable. In distinction from the channel casts, these casts reveal the large-scale trough cross-bedding viewed on a vertical section transverse to flow and the wavy boundary sets viewed on a vertical section parallel to flow. In these casts, cross-laminated series are up to 0.6 m thick (Plate II, fig. 6). In both cast types, accumulations of flattened mudstone pebbles are confined to boundaries between cross-laminated series.

In addition to casts, there are thin (0.1 to 0.5 m) beds of cross-laminated sandstone, which are traceable along the strike over a distance of several kilometers (Plate II, fig. 4). Their pinching-out in lateral direction has not been observed, although it is also possible that they represent deposits of wide (several kilometers across) distributary channels.

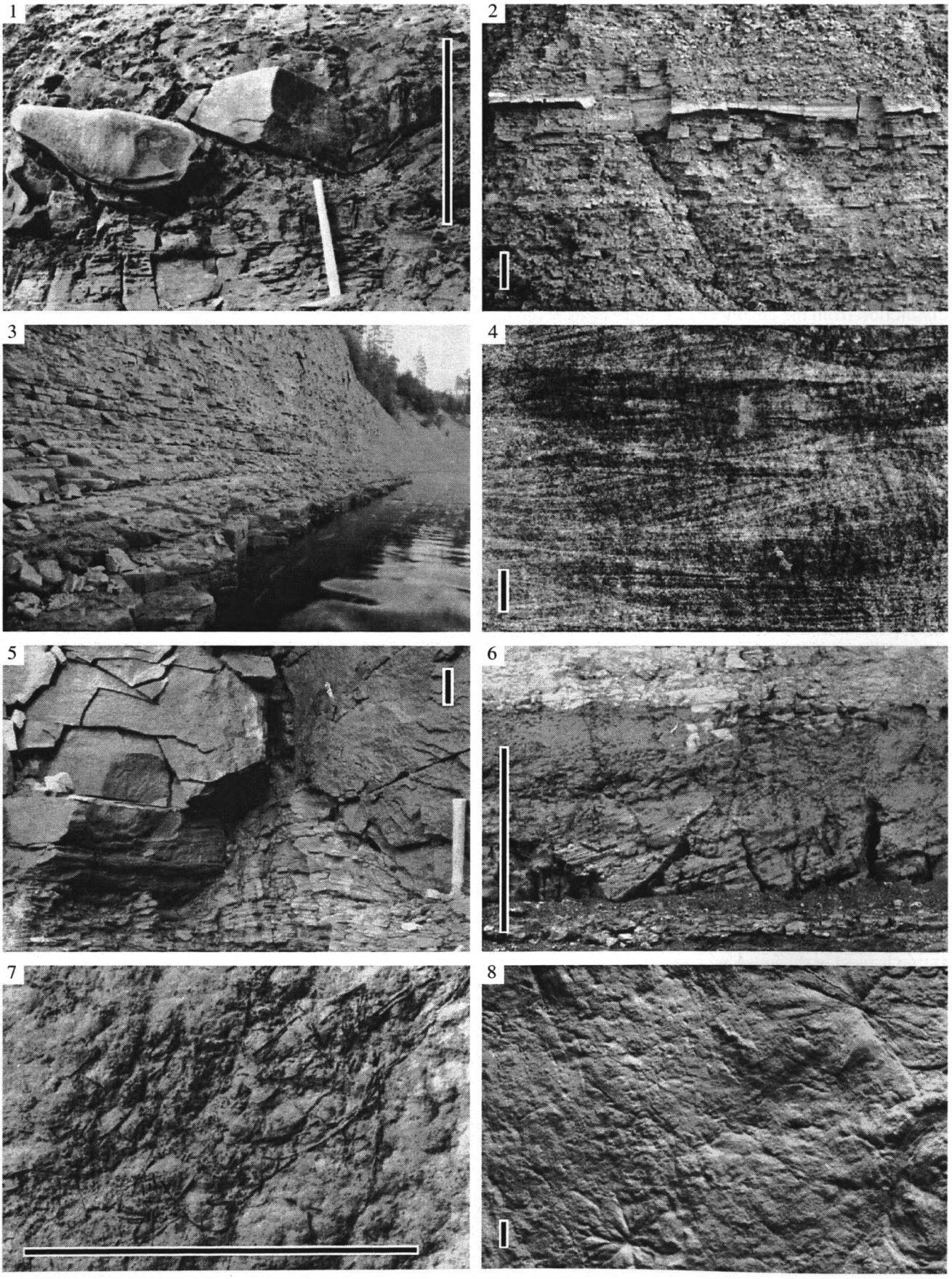
LITHOSTRATIGRAPHY

Exposures and borehole sections of sedimentary cover, which have been studied along the eastern margin of the Baltic Shield, elucidate important details of the Vendian Complex structure. In all the sections, the Vendian succession is divisible into large packages, each differing from the other in the content of sandy material and, sometimes, in coloration. These subdivisions are well traceable within a given facies zone only, but their regular alternation in the succession is persistent and can be used to distinguish the natural parageneses of lithologic types characterizing individual sedimentary cycles and their stratigraphic succession (Fig. 4) that is used in this work as a basis for subdivision and correlation. Sedimentary cycles recognizable in the Vendian succession range in scale from microcycles characterizing texture of separate lithologic types to mini-, meso-, and megacycles, which exemplify corresponding periods of sedimentation (Fig. 4).

Vendian deposits of the southeastern White Sea area have turned out to be of a more complex lithology and facies variability than it was thought before. For instance, a thick sandstone sequence terminating sections in the Onega Peninsula (Fig. 2) was erroneously interpreted as signifying the beginning of a transgressive cycle and correlated with the Mezen (Erga in this work) Formation (Aksenov, 1985; Stankovsky *et al.*, 1985). Recent data (Grazhdankin and Bronnikov, 1997) showed, however, that this sequence is coarsening upward, in contrast to fining-upward sediments of the Erga Formation, and terminates the corresponding sedimentary cycle. In addition, drilling results obtained in the Onega Peninsula changed views on stratigraphic position of cross-bedded gravelstone beds, which were originally interpreted as early transgressive deposits of the Ust-Pinega Formation (Solontsov *et al.*, 1970) and then attributed to basal Tamitsa Member of the latter (Stankovsky *et al.*, 1990) thereby representing marine ingression event (Fig. 1). New boreholes drilled in the Onega graben demonstrate that these gravelstones form progradation wedges in sandstone fill of that graben and pinch out toward the graben axial zone (Fig. 3). According to textural peculiarities, the host sandstones are similar to underlying Upper Riphean deposits. Consequently, the gravelstone wedges are of the same age, as one can judge from their genesis and structural position.

Admitting that priority and stability of stratigraphic nomenclature are important, I suggest to restore the formation names originally used by Kalberg in her reports on stratigraphy of the southeastern White Sea area, though stratigraphic ranges of the subdivisions must be changed according to new data (Fig. 1). In their new ranges, each of the Lyamtsa, Verkhovka, and Zimnie Gory formations on the one hand, has persistent lithologic-facies characterizations, and on the other hand, represent an integral succession. Kalberg suggested the names after investigation of known natural exposures known at that time and core sections recovered from boreholes Nenoksa and Arkhangelsk (unpublished report of 1950 by Kalberg and Frumkina). Stankovsky and his colleagues used the same names, when they divided the Ust-Pinega Formation into members, which are still used in publications and geological practice (Fig. 1). The term Erga Formation is introduced to

Plate I. Principal lithofacies of the Vendian Complex, southeastern White Sea area: (1) sandstone interbed with ripple cross-lamination, Borehole C18 (depth interval 179.0–182.0 m, core diameter 52 mm); (2) thin-bedded mudstone with volcanic ash interlayer (light colored), Lyamtsa Formation, Borehole C17 (depth interval 300.0–305.4 m, core diameter 52 mm); (3) interbedded siltstone and mudstone, Lyamtsa Formation, Borehole C11 (depth interval 121.0–126.0 m, core diameter 52 mm); (4) flute casts in interbedded siltstone and mudstone, Verkhovka Formation, Zimnie Gory site (scale bar 0.1 m); (5) siltstone interlayer with cross-lamination in interbedded siltstone and mudstone, Verkhovka Formation, Zimnie Gory site (scale bar 10 mm); (6) thin-bedded siltstone Zimnie Gory Formation, Zimnie Gory site (scale bar 10 mm); (7) flaggy stratification of alternating fine-grained sandstone beds and interbedded siltstone and mudstone, Verkhovka Formation, lower reaches of the Syuzma River (scale bar 0.1 m); (8) swing marks at the base of sandstone bed, Zimnie Gory Formation, Zimnie Gory site (scale bar 10 mm); (9) isolated bodies produced by syngenetic deformations in fine-laminated sandstone bed; Erga Formation, Zimnie Gory site (scale bar 1 m).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replace the Mezen Formation having a homonym. The term "Mezen Formation" was earlier introduced in the Upper Permian deposits of Mezen syneclise in 1969 as an equivalent of the synonymous horizon established by A.A. Malakhov in 1940 (see in Molin *et al.*, 1986).

Age of the Padun Formation has not been determined (Sivertseva and Stankovsky, 1982), and this subdivision is omitted from consideration in this work. It is admissible that the indicated formation includes Cambrian strata penetrated by boreholes in the study area (Popov and Gorjansky, 1994). The Vendian upper boundary provisionally corresponds to the base of a thick sandstone member (2.5 m) with planar lamination and conglomerate deposit. The boundary is observable in the Myandovo Canyon (Zolotitsa River middle reaches), where it corresponds to the base of Zolotitsa Beds of the Padun Formation (Stankovsky *et al.*, 1990).

Lyamtsa Formation has been established by Kalberg in coastal cliffs near the village Lyamtsa and in the borehole Arkhangelsk, depth interval of 533.9–435.9 m (Fig. 2), as a unit underlying the Arkhangelsk Formation of sandy deposits (the same borehole, depth interval of 435.9–372.5 m). Both formations represent an integral sedimentary megacycle of coarsening-upward deposits, and it is reasonable to consider them coupled under the name of Lyamtsa Formation. With this new range, the formation spans interval of the Lyamtsa and Arkhangelsk beds in the stratigraphic scheme by Stankovsky (Fig. 1).

The section exposed in coastal cliffs of the Lyamtsa River mouth can be retained as the formation stratotype (Fig. 3), because the general succession of deposits is established in the borehole drilled near the exposure in 1950, which penetrated the formation base at the depth of 77.4 m (Zoricheva, 1963). A candidate for hypostratotype is the depth interval of 214–54 m in the borehole S18 drilled in the right side of the Agma River valley by the Syuzma Geological Party in 1996. The formation thickness is rather persistent, equal to 160 m in the type section (Fig. 2).

The formation lower boundary corresponds to the base of gravelstone package, which displays graded bedding and is overlain by a member (12 to 18 m thick) of variegated thin-laminated mudstone, brown to gray in coloration and hosting the volcanic ash interlayers (Plate I, fig. 2). Toward the northeast (borehole Arkhangelsk), the mudstone grades into a variegated sequence of interbedded siltstone and mudstone (report of 1951 by A.I. Lebedintsev), the mudstone component of which is prevailing (Fig. 2). The lower boundary stratotype is suggested to be in the borehole 18 (Fig. 2) at the depth level of 214 m corresponding to the base of thin (0.4 m) package of light gray gravelstone.

A uniform rhythmical interbedding of thin variegated siltstone and mudstone beds characteristic of the formation (Plate I, fig. 3) is interrupted by progradation wedges enclosing interlayers (20 to 140 mm) of gray sandstone with ripple cross-lamination (Plate I, fig. 1). Spatial position of wedges is reconstructed based on layer-by-layer correlation of borehole sections (Figs. 2 and 3). This structural peculiarity offers a possibility to divide formation into three members representing a succession of sedimentary mesocycles, the thickness of which progressively increases upward in the section (24, 50, and 86 m, respectively, in the borehole section C18; Fig. 2). The progradation wedges are situated near the shield slope, and away from the latter, the cyclic structure of the formation is not as distinct (Fig. 3). Each mesocycle represents a regressive succession, the abundance and thickness of sandstone interlayers in which grow upward. The total thickness of sandstone interlayers increases in each successive mesocycle (Fig. 2).

The interbedding siltstones and mudstones bear phytoleims of vendotaenian algae, organic-walled remains *Beltanelloides sorichevae* Sokolov, and wrinkled tubular sheaths resembling *Sabellidites*. Swing marks widespread on sole planes of rare sandstone interlayers are broadly arched or bimodal in morphology. Remains of Ediacara-type biota are rare in the Lyamtsa Formation and occur mostly in the upper sandy portion of the upper mesocycle (Fig. 4).

Verkhovka Formation was established by Kalberg in the borehole Arkhangelsk (depth interval of 372.5–299.6 m; Fig. 2). Its stratotype has not been indicated, but the name derives from the eponymous river valley, where (Kalberg, 1940) she studied outcrops. The boundary with overlying Zimnie Gory Formation was placed at the base of a sandstone member (depth level of 299.6 m in the borehole Arkhangelsk; Fig. 2). Exposures studied in the Onega Peninsula show that this member is terminal one in a thick (140 m) regressive mesocycle, and it exhibits gradual transition into the underlying deposits of the Verkhovka Formation (Figs. 3 and 4). Therefore, the revised formation range

Plate II. Principal lithofacies of the Vendian Complex, southeastern White Sea area: (1) isolated gutter casts at the boundary between the Zimnie Gory and Erga formations, Zimnie Gory site (scale bar 1m): (2) sandstone interlayer with hummocky stratification in interbedded siltstone and mudstone, Verkhovka Formation, Letnii Coast of the White Sea (scale bar 0.1 m); (3) thick sandstone member with planar lamination, Verkhovka Formation, lover reaches of the Solza River (apparent thickness 15 m); (4) multistoried cross-lamination, Zimnie Gory Formation, Zimnie Gory site (scale bar 10 mm); (5) two waterways recorded at the base of sandstone channel cast, Erga Formation, Zimnie Gory site (scale bar 0.1 m); (6) longitudinal section of a channel cast with basal cross-laminated sandstone grading upward into sandstone with planar lamination, Erga Formation, middle reaches of the Zolotitsa River (scale bar 1 m); (7) pyritized remains of algal filaments on a shagreen under surface of thin-bedded sandstone, Erga Formation, Zimnie Gory site (scale bar 10 mm); (8) deformation marks at the shagreen under surface of siltstone bed in the interbedded siltstone-mudstone sequence, Verkhovka Formation, Zimnie Gory site (scale bar 10 mm).

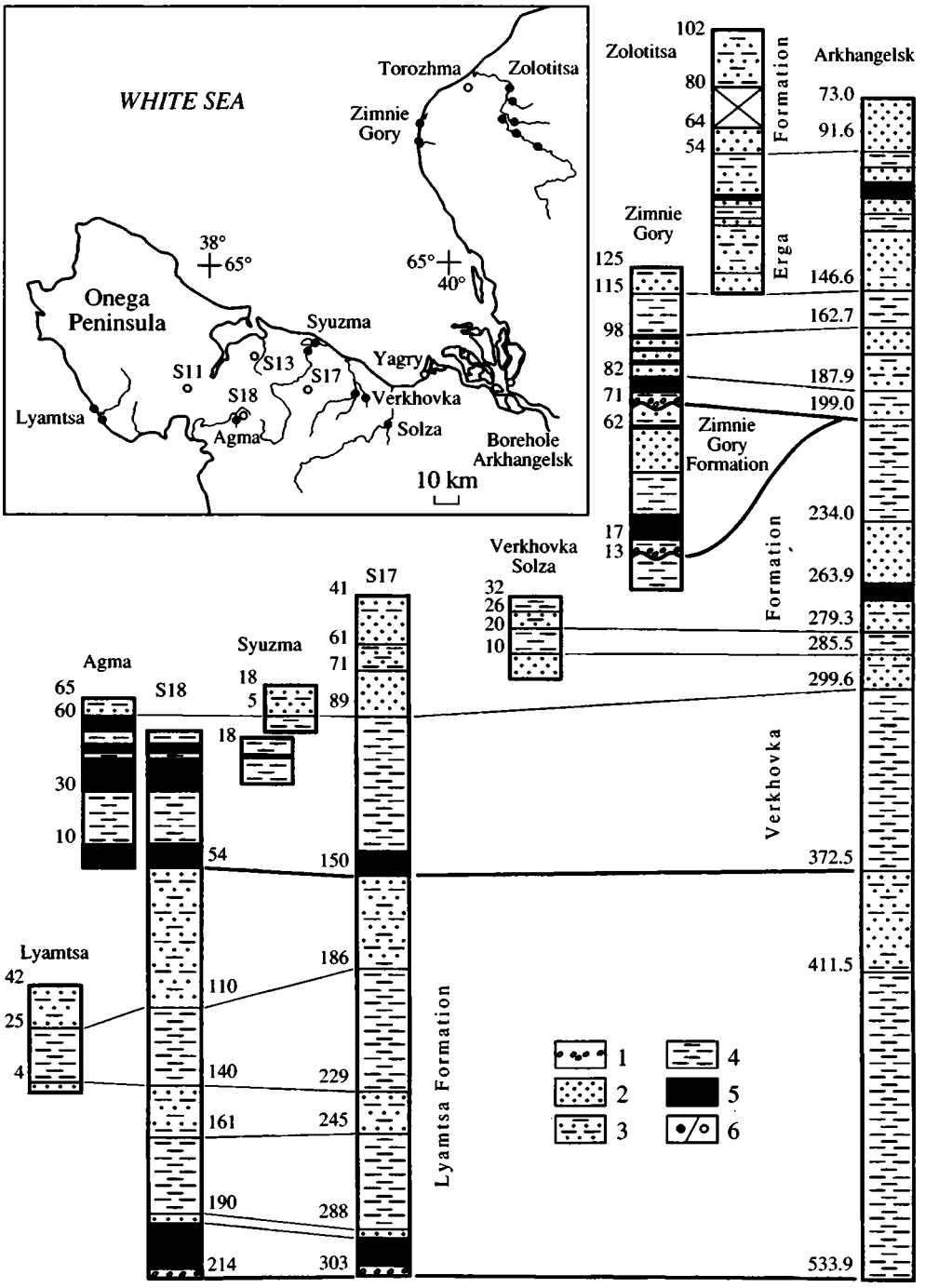

Fig. 2. Geographic localities and lithostratigraphy of Vendian sections studied in the southeastern White Sea area: (1) gravelstone; (2) sandstone; (3) interbedded sandstone, siltstone, and mudstone; (4) interbedded siltstone and mudstone; (5) mudstone; (6) exposure/borehole (section Arkhangelsk is plotted based on unpublished data of A.I. Lebedintsev, 1951).

is greater than the original one and includes a part of Zimnie Gory Formation in the former understanding of Kalberg (Fig. 1). The new range spans the interval of Verkhovka and Syuzma beds in the scheme by Stankovsky (Fig. 1). The depth interval of 109.8–81.0 m in the borehole Yagry, which has been attributed by Kalberg to the transitional Yagry Formation (Fig. 3), is well correlative with the middle portion of the Verkhovka Formation. Accordingly, I consider the term "Yagry Formation" as a synonym of the latter. The most complete section of the Verkhovka Formation (170 m; Fig. 2) has been penetrated by the borehole Arkhangelsk (Fig. 2), the core samples of which are lost.

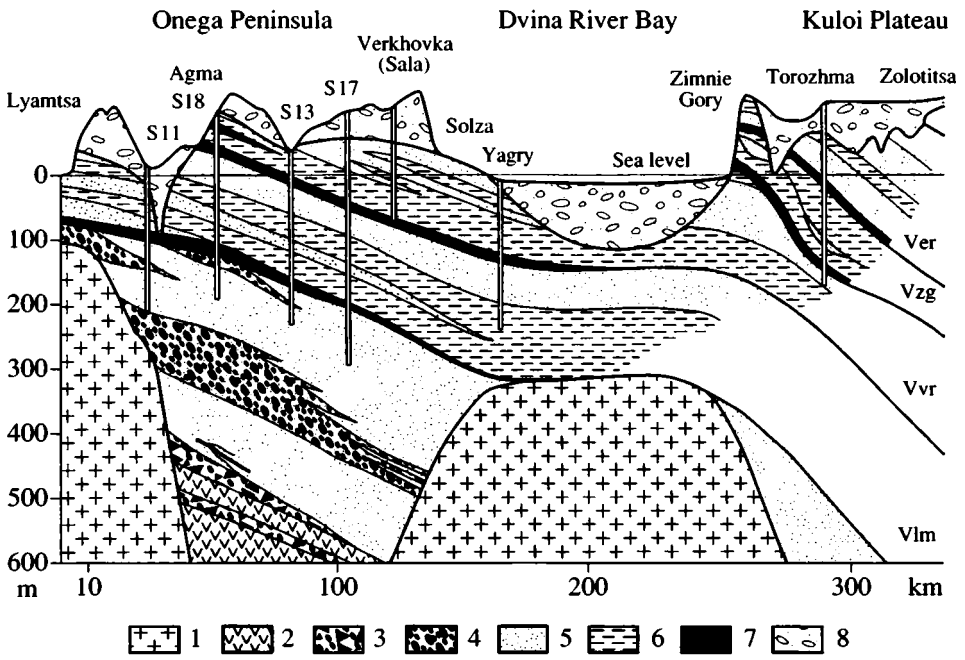

Fig. 3. Suggested architecture of the Vendian and underlying deposits in the southeastern White Sea area: (1) crystalline basement; (2) basalt; (3) volcanic breccia; (4) gravelstone; (5) interbedded sandstone, siltstone, and mudstone; (6) interbedded siltstone, and mudstone; (7) mudstone; (8) Quaternary deposits. Letter symbols denote Lyamtsa (Vlm), Verkhovka (Vvr), Zimnie Gory (Vzg), and Erga (Ver) formations. Borehole sections are described by Kalberg (1937; Yagry and Sala), Zoricheva (1963, Lyamtsa), Stankovsky *et al.* (1990, Torozhma).

The Verkhovka Formation represents a regressive megacycle. Its lower boundary is established at the level of abrupt transition from interbedding gray sandstone, siltstone, and mudstone of the Lyamtsa Formation to overlying member of maroon fine-laminated mudstone with volcanic ash beds. The lower boundary stratotype is suggested to be fixed at the base of maroon mudstone member (10 m thick) viewed in exposure of the Agma Creek mouth site. U–Pb age of zircons from the ash bed (12 mm) located in that exposure 0.5 m above the formation base is 558 ± 1 Ma (Martin *et al.*, 2000).

The megacycle consists of two mesocycles, the lower of which is exposed in outcrops of the Onega Peninsula (Fig. 2), where its lower half viewed in the Agma Creek outcrops (left tributary of the Nizhma River) represents the alternation of members (4 to 30 m thick) composed either of variegated fine-laminated mudstone, or of interbedded greenish gray siltstone and mudstone. The latter lithologic type also encloses thin (10 to 15 cm) lenticular sandstone interbeds displaying hummocky stratification (Plate II, fig. 2) and small (30 to 50 mm) gutter casts. Eastward, the basal mudstone member grades into facies of interbedded mudstone and siltstone. Close to the base of the lower mesocycle, there are widespread sulfide, gypsum, and carbonate concretions viewed in outcrops of the Agma, Syuzma, and Letnii Coast localities and observed in the borehole S17, Sala, and Yagry. The flattened (disk-like) shape of concretions and cone-in-cone structures characteristic of carbonate ones imply that they were formed in partially compacted sediments.

The upper sandy interval of the lower mesocycle is of a complex structure. It has been penetrated by borehole S17 (depth range of 89–41 m) and is fragmentary exposed in the Syuzma, Agma, Verkhovka, and Solza river valleys (Figs. 2 and 3). In western sections, the interval is composed of sandstone beds striking NE–SW and well studied in a small outcrop located 5 km upstream of the Syuzma River mouth (Fedonkin, 1981). The basal, purple-gray thin-bedded siltstone encloses here lenticular gutter casts (0.3–0.4 to 1.2 m wide) composed of greenish gray, fine- and cross-laminated sandstone. The main stratified sequence (Plate I, fig. 7) is composed of lenticular sandstone beds (0.1 to 0.4 m thick), which display either thin horizontal bedding, or multistoried cross-lamination, and diverse erosion marks at their soles. Unimodal swing marks with an acute arching angle have been viewed on one of the soles. Thicker (10 m) members of planar-laminated sandstone are rather persistent along the strike, traceable for a distance as long as dozens of kilometers (Plate II, fig. 3).

Composition and structure of the interval under consideration change considerably across the beds' strike, toward the southeastern Verkhovka and Solza localities, where the massive sandstone beds either pinch out, or grade into layers of interbedded sandstone, siltstone, and mudstone so that the total proportion of sandy

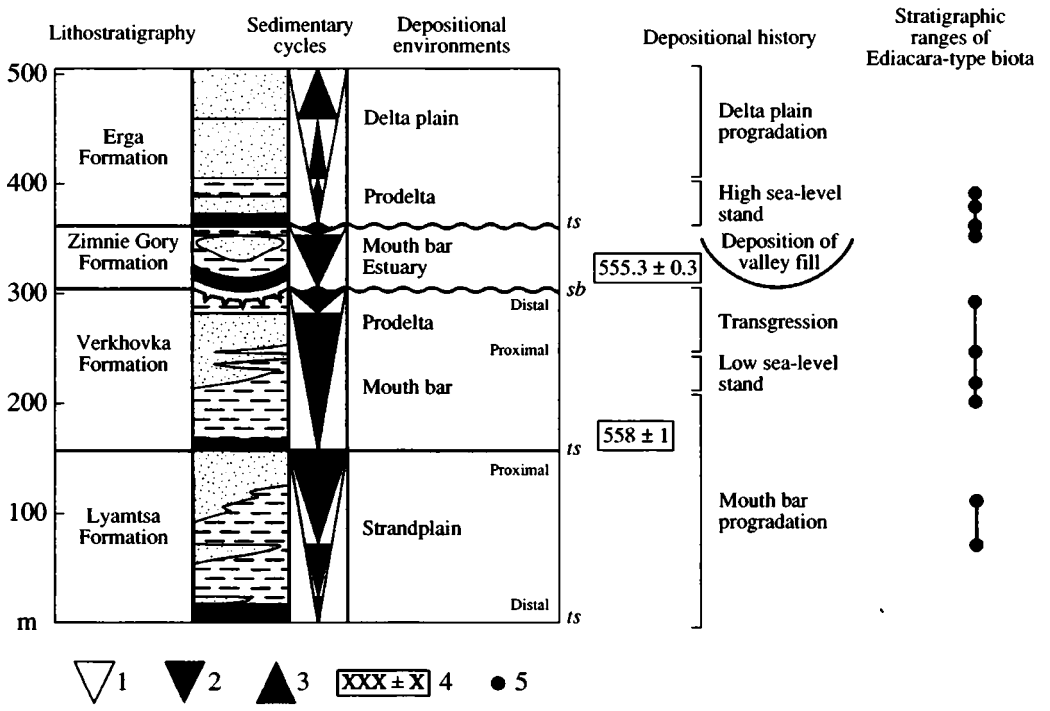

Fig. 4. Lithostratigraphy, depositional environments and formation history of Vendian deposits in the southeastern White Sea area: (1) regressive megacycle; (2) regressive mesocycle; (3) transgressive mesocycle; (4) radiometric age, Ma; (5) occurrence of diverse Ediacara-type biota; (*sb*) sequence boundary; (*ts*) transgressive boundary (other symbols as in Fig. 3).

material is lower here (Fig. 3). Characteristic of this interval, lenticular thin-bedded sandstone beds (0.3 to 0.5 m thick) have erosional soles and strike NE–SW, but the strata are locally deformed into folds and various slump structures. The deformed units occur between the undistorted ones, and the general style of deformations is irregular, changing from site to site (Grazhdankin and Bronnikov, 1997).

In outcrops of the Zimnie Gory area, an interval (0–13 m; Fig. 2 and Plate I, figs. 4 and 5) of greenish gray interbedded siltstone and mudstone at the base of the section encloses interlayers of cross-laminated siltstone. Higher in the sections, there are sandstone interbeds (5 to 10 cm thick) with ripple cross-lamination. It is difficult to define the exact stratigraphic position of these rocks, because the area under consideration is situated far away from the Onega Peninsula, deep boreholes have never been drilled here, and the contact with overlying strata is erosional (Figs. 2–4). In this paper, they are correlated with the interval of 234.0–199.0 m of borehole Arkhangelsk (unpublished report of 1951 by Lebedintsev) and attributed to the upper mesocycle of the Verkhovka Formation (Fig. 2).

Ediacara-type remains are irregularly distributed in the described formation. Rare fossils are found near the base of the lower mesocycle and are associated with the drag and deformation marks on basal surfaces of sandstone beds with hummocky stratification. Pyritized remains of *Sabellidites*-like tubular morphotypes, which have been discovered by Kalberg, are known from the fine-laminated mudstone. Sandstone beds exposed in the Agma and Syuzma river valleys yield a peculiar assemblage of three-dimensional Ediacara-type fossils (Fedonkin, 1981). A diverse assemblages of fossils are also known from higher strata of interbedded siltstone and mudstone with sandstone interlayers, which are exposed in the Verkhovka and Solza river valleys (Grazhdankin and Bronnikov, 1997), and from the upper mesocycle exposed in the Zimnie Gory cliffs (Fedonkin, 1981; Fig. 4).

Sections, which are exposed near the base of these cliffs and in the Agma, Syuzma, Verkhovka, and Solza river valleys, comprise almost complete succession of the formation and can be accepted for the composite lectostratotype of the latter.

Zimnie Gory Formation has been distinguished by Kalberg in synonymous coastal cliffs and correlated with the depth interval from 299.6 to 146.6 m of the borehole Arkhangelsk (Fig. 2). In this work, its range is changed based on new data. In the new understanding, the formation is bounded by the lower and upper erosion surfaces and spans the interval attributed by Stankovsky to the Vaizitsa and Zimnie Gory beds (Fig. 1). In the northeast of study area, deposits of the formation are distributed irregularly. Being undetected in the borehole Arkhangelsk, the formation is 58 m thick in the Zimnie Gory section (interval from 13 to 71 m; Fig. 2) and 125 m thick in the borehole Torozhma

(depth interval of 261.5–134.9 m; Fig. 3) that was drilled by the Tova Geological Party 26 km northeastward of the Zimnegorskii lighthouse in 1982.

The formation lower boundary is established at the level of abrupt transition from greenish gray interbedded siltstone and mudstone of the Verkhovka Formation to variegated deposits with beds and members of maroon thin-laminated mudstone. Above this erosional boundary, there are lenticular beds (10 to 20 cm thick) of gravelstone and conglomerate. The overlying interval of variegated rocks is composed of sandstone, siltstone, and mudstone exhibiting the wavy and lenticular interbedding (interval from 13 to 17 m in the Zimnie Gory section; Fig. 2). Characteristic of lenticular siltstone and sandstone beds is multidirectional cross-lamination. These beds with erosion soles and wavy upper boundaries are thicker (2 to 6 cm) than the mudstone beds. In addition, the interbedded member encloses isolated gutter casts (15 to 20 cm thick), which are composed of interbedded sandstone and mudstone.

The upper boundary of the described unit is erosional, with scours locally as deep as 4 m. The overlying member (10 m thick) is composed of variegated, maroon to bluish gray thin-laminated mudstone with volcanic ash intercalations. The basal ash bed is conformable to uneven surface of underlying deposits. Ash beds are up to 10 cm thick in scours and pinch out approaching walls of the latter. U-Pb age of zircons from one of the interlayers, which has been sampled 2 m above the formation base in an outcrop near the Medvizhii Creek mouth is 555.3 ± 0.3 Ma (Martin *et al.*, 2000).

A considerable portion of the formation (up to 30 m thick) is composed of purple-gray thin-bedded siltstone (Plate I, fig. 6) with lenticular sandstone intercalations displaying ripple cross-lamination and swing marks on the soles (Plate I, fig. 8). Being rather uniform in the type section, the formation has variable structure in southern exposures (between the Zimnegorskii lighthouse and Erga Creek). Its upper interval (18 m) is represented here by gray sandstone beds of diverse facies affinity (Igolkina, 1959). The sandstone members (up to 2 m thick) comprising rocks with thin planar- and cross-lamination alternate with thin-laminated siltstone members of a comparable thickness, and the whole interval is of a rhythmic structure. The normal succession of beds is locally disturbed by slump structures. Ripple cross-lamination is not characteristic of these rocks. The interval under consideration corresponds to a regressive succession: thin sandstone beds with multistoried cross-lamination are typical of its lower part, whereas the upper one is composed of thick sandstone beds displaying thin planar lamination.

The upper part of the formation (interval from 62 to 71 m in the Zimnie Gory section; Fig. 2) composes a regressive sequence beginning with a persistent basal unit (3 to 4 m thick) of dark gray thin-laminated mudstone with sapropel-like films. The unit has gradual transition to overlying greenish gray siltstone and mudstone, thin alternating beds of which enclose lenses of gray thin-bedded sandstone. This sequence is comparable with the member of interbedded siltstone and mudstone that has been penetrated by the borehole Torozhma (depth interval of 208.3–134.9 m; Fig. 3) and represents a similar regressive sequence (report of 1985 by Stankovsky and his colleagues).

Rare Ediacara-type fossils are confined in the formation section to the upper sequence of interbedded siltstone and mudstone. In addition, the clay unit near the formation base has yielded phytoleims of algae (M.V. Leonov, personal communication, 2001) and *Beltanelloides* impressions.

The exposure of Vendian rocks near the Elovyi Creek mouth is suggested to be the lectostratotype of the Zimnie Gory Formation. The lower boundary stratotype corresponds here to the under surface of maroon fine-laminated mudstone bed (1 m thick) with basal gravelstone lenses.

Erga Formation introduced instead of the Mezen Formation spans the interval of Egra and Mela beds in the scheme by Stankovsky (Fig. 1). Like the synonymous beds, it has got the name after the eponymous creek (Zimnie Gory area), because characteristic rocks of the formation are exposed near the mouth of the latter. Typical of the formation representing a coarsening-upward sedimentary megacycle (150 m thick) is a complex interbedding of variegated sandstone, siltstone, and mudstone. The lower boundary of the subdivision is characterized by erosional scours (1 to 5 m deep), the fine-grained sandstone fill of which encloses pebbles occurring in the imbricated position within the matrix. In the stratigraphic succession the boundary corresponds to the first appearance of yellowish gray sandstone with gravel and pebbles.

Three transgressive mesocycles (45 to 55 m thick) are distinguishable in the formation section. The lower one exposed in coastal cliffs of the Zimnie Gory area (interval from 71 to 115 m; Fig. 2) consists of three persistent members. In its lower part, the lower member (interval from 71 to 82 m) is composed of interbedded sandstone, siltstone, and mudstone (2 to 3 m), which enclose large gutter casts (Plate II, fig. 1). Ripple cross-lamination is characteristic of their basal beds. The whole member represents a fining-upward sequence, because the indicated rocks have gradual transition to overlying gray mudstone (8 to 9 m) that becomes variegated, from brown to gray, in the uppermost portion.

The middle member (interval from 82 to 98 m) is divisible into persistent cyclothems 5 to 6 m thick (Fig. 2). At the base of each cyclothem, there are gutter casts and lenticular beds (up to 0.5 m thick) of gray sandstone with ripple cross-lamination. They grade upward into thin-interbedded siltstone and mudstone with sandstone interlayers followed by fine-laminated mudstone. This succession is locally deformed into folds and various slump structures, some of which have

been erroneously interpreted as gutter casts (Grazhdankin and Ivantsov, 1996, Fig. 2). In contrast to slump pillows, the gutter casts are elongated in shape and represent erosional incisions.

The upper member (interval from 98 to 115 m) reveals thin wavy interbedding of greenish gray sandstone, siltstone, and mudstone. Sandstone beds with characteristic ripple cross-lamination enclose accumulations of flattened mudstone pebbles. In addition, there are isolated gutter casts (10 to 15 cm) and peculiar lenticular beds (up to 1.8 m) of planar- and cross-laminated sandstone (casts of distributary channels, Plate II, fig. 5), which occupy several levels in the interbedded sequence.

Beginning from the middle mesocycle (the level of 115 m in the Zimnie Gory section; Fig. 2), sandstone beds of the Erga Formation are colored red, bearing large siderite concretions, whereas the interbedded siltstone and mudstone become variegated (purple, bright-brown, yellow, to orange). The same changes in coloration are recorded at the depth level of 146.6 m in the borehole Arkhangelsk (report of 1951 by Lebedintsev; Fig. 2). The middle mesocycle is exposed in the Zolotitsa (Zimnyaya) River lower riches and along the Temnyi Creek, the right tributary of that river (interval from 0 to 54 m of the Zolotitsa section, Fig. 2). Basal beds (2 m) of the mesocycle are composed of sandstone with planar bedding and current-formed parting lineation. The overlying part is represented by thin-interbedded sandstone, siltstone, and mudstone, which enclose lenticular beds (up to 2.5 m) representing gutter casts of medium- to coarse-grained sandstone composition, which reveal planar and cross lamination and enclose flattened mudstone pebbles (Plate II, fig. 6). In the upper part of the mesocycle, there are beds (up to 1.2 m) and members (up to 6 m) of dark brown fine-laminated mudstone. Ripple cross-lamination and ripple marks are characteristic of rocks constituting the mesocycle.

The upper mesocycle is exposed further upstream along the Zolotitsa River and its right tributary, the Kamennyi Creek (interval from 54 to 102 m of the Zolotitsa section, Fig. 2). It corresponds to the distinct fining-upward sequence and resembles the middle mesocycle.

The assemblage of Ediacara-type remains that is the most diverse among others known from the southeastern White Sea area is confined to the lower part of the Erga Formation, being especially characteristic of the middle member of the lower mesocycle (Grazhdankin and Ivantsov, 1996; Ivantsov, 1999). Above this interval, fossils have not been encountered. In the Erga Formation, there is recorded a transition from the Redkino to Kotlin assemblage of organic-walled microfossils (Ragozina and Sivertseva, 1990). Microfossils known from the Lyamtsa, Verkhovka, Zimnie Gory, and lower Erga deposits are characteristic of the Redkino assemblage (Sivertseva and Stankovsky, 1982; Ragozina and Sivertseva, 1990; Stankovsky *et al.*, 1990). Microfossils *Pomoria rhomboidales* (Sivertseva), which have been described from the upper member of lower mesocycle of the Erga Formation (depth level of 57 m in the borehole Torozhma, Fig. 3), are widespread in the upper Kotlin beds of the Moscow syneclise (Ragozina and Sivertseva, 1990; Kuz'menko and Burzin, 1996).

The formation neostratotype is suggested to be corresponding to a sequence of sections exposed in the Zimnie Gory area (between the Zimnegorskii lighthouse in the north and Erga Creek mouth in the south) and along the Zolotitsa River, which characterize almost completely the rock succession described above. The candidate for the lower boundary stratotype is the under surface of member composed of interbedded siltstone and mudstone, which encloses gutter casts of sandstone composition and is exposed in outcrops of the Erga Creek mouth.

Cyclicity

Mesocycles ranging in thickness from 40 to 80 m are most easily distinguishable in the studied sections and can be divided in two types with different lithologies and bed successions. Basal portions of the regressive mesocycles are composed of fine-grained sediments exclusively (e.g., of thin-laminated mudstone). These sediments are overlain by interbedded siltstone and mudstone sequence that encloses sandstone interlayers with thin planar- or ripple cross-lamination, the abundance and thickness of which increase progressively upward. Mesocycles of this type are established in the Lyamtsa, Verkhovka, and Zimnie Gory formations (Fig. 4). In the transgressive mesocycles, basal sandstone beds reveal planar bedding near the base, fine- and cross-lamination in the middle interval, and ripple cross-lamination in the upper one. They are overlain by a fining upward interbedded siltstone and mudstone sequence. The transgressive mesocycles are established in the Erga Formation. Thus, the boundary between the latter and the underlying Zimnie Gory Formation corresponds to the sudden change in morphology of sedimentary mesocycles of the Lyamtsa, Verkhovka, and Zimnie Gory formations (Grazhdankin and Bronnikov, 1997; Fig. 4).

Mesocycles are divisible in lesser sedimentary cycles (10 to 20 m thick), which can be distinguished, for instance, in succession of the Verkhovka Formation rocks (Grazhdankin and Bronnikov, 1997), though it is difficult to trace them for a long distance, because the outcrops are not so well exposed here. Minicycles are clearly recognizable in the lower part of the Erga Formation, and this interval of the Zimnie Gory section is divisible into corresponding persistent members. The latter are composed of elementary cyclic units (5 to 6 m thick) resembling cyclothems of deltaic deposits. These cyclothems of fining-upward structure have been traced along the Zimnii Coast of the White Sea for a distance of 35 km, from the Zimnie Gory site to the Zolotitsa

River mouth. They are of persistent thickness and lithologic composition.

Combinations of mesocycles represent larger megacycles 120 to 160 m thick, which are traceable over great areas. The Vendian succession of the study region is divided into four megacycles of this kind, which correspond to the formation described above and are in general of the coarsening-upward structure (Fig. 4).

DEPOSITIONAL ENVIRONMENTS

Data presented above enable reconstruction of depositional environments during accumulation of the Vendian Complex.

Depositional environment of the Lyamtsa Formation. It is likely that most distal deposits of the formation are represented by fine-laminated mudstone. Their undisturbed fine lamination (Plate I, fig. 2) points to precipitation of suspended sedimentary material in low energy conditions. Like in the present-day seas (Oertel and Dunstan, 1981; Holmes, 1982; Sahl *et al.*, 1987), the suspended material could be formed in offshore settings and, still floating owing to difference in temperature and salinity, it could be sorted by coastal currents and during strong storm events and then transported away from the shore into depositional settings. The upper silty and muddy layers of bottom sediments could be recurrently suspended under influence of storm waves and, being captured by bypassed flows, the material was gradually transported toward a zone, where the oscillation wave rate was lower. The progressive sorting in wave flows (Swift *et al.*, 1991) resulted in deposition of coarser silty particles within the proximal zone, where the thin-interbedded siltstone-mudstone sequence (Plate I, fig. 3) characterizing the greater part of the formation section was formed. Thin sandstone interlayers with ripple cross-lamination (Plate I, fig. 1), which occur close to the shield slope, the widely arcuate or bimodal swing marks on their soles, and multidirectional current-formed parting lineation are important arguments in favor of the wave mechanism of clastic material transport. Fine-laminated mudstone accumulated in distal settings inaccessible for the storm wave impact.

Thus, the greater portion of the formation was deposited in strandplains above the storm wave base. The perfect sorting and indications of wave reworking allow me to interpret the interbedded siltstone-mudstone sequence with sandstone interlayers as storm deposits. Widespread algae phytoleims and swing marks on the interlayers' soles imply that strandplains were situated within the photic zone and inhabited by algae. Progradation wedges of sandstone reflect most likely a higher rate of wave reworking and sediment transport on shoals, and the progradation of strandplains away from the eastern slope of the Baltic Shield (Figs. 2 and 3).

Depositional environment of the Verkhovka Formation. The base of the formation marks sea level rise and abrupt environmental changes. Strandplains of the terminal Lyamtsa time turned into relatively calm settings of predominantly clayey sedimentation (boreholes 17 and 18, and the Agma Creek section; Fig. 3), and zone of wave flow influence shifted to the northeast, where deposition of interbedded siltstone-mudstone sequence was again in progress (Arkhangelsk).

The further development was associated with gradual return of offshore settings with storm-wave sedimentation regime, as it is evident from a thick (50 m) interbedded siltstone-mudstone sequence occupying the lower formation part in the Onega Peninsula (Fig. 3). Rare and thin sandstone interlayers in these deposits have hummocky stratification (Plate II, fig. 2) that used to be regarded as indicative of wave activity. However, the influence of oscillating flow is not the only factor responsible for origin of this lithologic type (Allen, 1985), because progressive gravity waves should interact in this case with a unidirectional turbid flow (Harms *et al.*, 1982; Walker *et al.*, 1983; Nottvedt and Kreisa, 1987). The hummocky cross-stratification in sandstone beds of the Verkhovka Formation is likely a result of such an interaction. Small isolated gutter casts characteristic of deposits under consideration also suggest a considerable contribution of turbid flow activity to the sandy material transport.

The stratified sandstone package (Plate I, fig. 7) in the middle formation part is traceable over a distance of 40 km, from lower reaches of the Syuzma River to the Agma Creek (Fig. 2). Its peculiar features, such as abrupt lower boundaries of the beds with current-formed erosion marks, combination of multistoried cross-lamination with thin horizontal bedding, and gutter casts, imply that the beds were deposited from turbid flows under conditions of intense influx of terrigenous material. Measurements of cross-lamination suggest a unimodal distribution of paleocurrent directions from the northeast (230°, n = 10). This is consistent with the unimodal character of swing marks and with the trend of flute casts (245°, n = 12) and of the package itself. The planar bedding characteristic of thick sandstone members seems to be a result of sand discharge from relatively rapid flows. This is evident from current-formed parting lineation that represents a response of a sand bed to flow configurations arising during boundary-layer streaks, provided the long-term regime and relatively high constant rate of sediment transport. According to all these data, the sandstone beds are interpreted as fluviomarine deposits.

In the southeastern White Sea area, western boundaries of present-day distribution area of the fluviomarine deposits are erosional. In the east, these deposits, predominantly sandstones with diverse slump structures, form thick (6 to 15 m) wedges oriented across the main direction of paleocurrents (Grazhdankin and Bronnikov, 1997). By all appearances, the quick accu-

mulation of sand caused a slight seafloor sloping that destabilized sediments. In the host sequence, sandstone beds with ripple cross-lamination are intercalated with members of interbedded siltstone and mudstone, and these rocks are interpreted as sediments deposited in a zone of wave agitation and current activity (Grazhdankin and Bronnikov, 1997). Lenticular beds of thin-bedded sandstone with erosional soles, which are also characteristic of this area and extend along the main direction of paleocurrents, likely represent casts of submarine channels.

Thus, the lower mesocycle of the Verkhovka Formation was formed in an offshore area in distributary-mouth bar setting, where a river flow interacted with rough sea. The thick sequence of fluviomarine deposits represents the delta mouth bar extending in direction of turbid flows. Slope deposits accumulated in prodelta settings of the mouth bar flanks. The flows episodically penetrated in these settings and locally scoured channels in the bottom, though in general they were of a low density. Decelerating in the prodelta settings, flows discharged the suspended material that was subsequently reworked by wave flows, which created the widespread cross-stratification.

The upper mesocycle of the Verkhovka Formation is composed of thin-interbedded siltstone and mudstone deposited under conditions of progressive sorting of bottom sediments by wave flows. This is evident from local erosion marks recorded in deposits (Plate I, fig. 4), from cross-lamination of siltstone interbeds (Plate I, fig. 5), and from sandstone intercalations with ripple cross-stratification, which enclose flat mudstone pebbles. According to these data, the siltstone-mudstone sequence (13 m thick) of the upper mesocycle represents distal deposits of prodelta indicative of a relative sea-level rise in the late Verkhovka time. The middle interval of fluviomarine deposits consequently corresponds to the lowest position of sea level in the Lyamtsa-Verkhovka succession (Fig. 4).

Depositional environment of the Zimnie Gory Formation. The formation lower boundary outlines, as is suggested, the gently sloping bottom relief of a submarine valley that was dozens of meters deep and dozens of kilometers wide (Fig. 3). Basal conglomerate infills erosional scours in underlying sediments and imply the erosional origin of the boundary.

Desiccation marks have not been observed in the studied formation sections. Basal conglomerates are overlain by mudstone deposited in relatively low energy conditions and bearing abundant algae phytoleims. Interbedded sandstone, siltstone, and mudstone occurring higher in the section can be attributed to estuarine sediments of the relatively shallow-water genesis based on their lithologic peculiarities and abundant plant detritus characteristic of the rocks.

The thick stratified package of thin-, planar-, and cross-bedded sandstones represents an important characteristic of the formation elucidating the depositional environment. In the northeast, these sandstone beds extend for a distance of 30 km between the Torozhma and Zimnie Gory sites. Distinct flute casts, multistoried cross-stratification (Plate II, fig. 4), and widespread syngenetic deformations, including deformed cross-stratification, imply fluviomarine origin of the rocks (Swift *et al.*, 1991). Cross stratification suggests paleocurrents from the northeast (210°, n = 9). Across the beds' strike, the composition and structure of sandstone member change. In the northwest (northern Zimnie Gory sections), sandstone beds pinch out, and the member is composed of thin-laminated siltstone strata (Plate I, fig. 6). The well-sorted material of the latter, interlayers with ripple cross-lamination, and broadly arcuate swing marks (Plate I, fig. 8) are indicative of sedimentation settings affected by storm hydrodynamics. According to all these data, the sandstone member is interpreted as deposits of the axial zone of a mouth bar.

The lateral facies succession of the Zimnie Gory Formation therefore reflects gradual progradation of an estuary-mouth bar sedimentation system. The system developed from an incipient submarine valley incised into underlying sediments (Fig. 4). Local erosion is also recorded at the top of estuarine deposits, but this level marks the abrupt transition from estuarine to fluviomarine sedimentation at an earlier stage of progradation. In the upper formation part, there is persistent member composed of fine-laminated mudstone grading into interbedded siltstone and mudstone with sandstone intercalations. These rocks imply that the mouth bar settings of the terminal Zimnie Gory phase suddenly turned into prodelta settings.

Depositional environment of the Erga Formation. Cyclothems in the lower formation part reveal a characteristic sedimentary succession that is important for understanding of their depositional environment. The basal beds and isolated gutter casts of cyclothems are composed of thin-bedded sandstone. They often enclose flat mudstone pebbles concentrated near soles with occasional erosion marks. Thin horizontal and graded bedding of the rocks suggests that they represent a discharge product of turbid flows (Grazhdankin and Ivantsov, 1996). A slight seafloor sloping in response to quick accumulation of sand triggered development of syngenetic slump deformations (Plate I, fig. 9), and the rocks can be accordingly regarded as deposits of distributary delta channels. The thick sandstone beds grade upward into interbedded sandstone, siltstone, and mudstone. The ripple cross-lamination and symmetrical ripple marks widespread in the interbedded member are indicative of sedimentation under influence of weak wave flows. The overlying member of interbedded siltstone and mudstone corresponds to strandplain storm deposits. Finally, the topmost member of cyclothems is composed of fine-laminated mudstone deposited in low energy conditions.

Gutter casts characteristic of proximal prodelta facies (Plate II, fig. 1) suggest a zone of strong bypass-

ing turbid flows that scoured gutters in seafloor. The gutters became immediately cast by coarse sediment partially discharged from turbid flows, which became more buoyant, ascended as plumes leaving behind the gutter casts, and transported the rest of suspended material to more distal settings.

Distributary and overflow channels filled with sediments are widespread in the upper member of lower mesocycle. Their soles with deep erosion marks and sandstone with planar lamination deposited in deepest parts (Plate II, fig. 5) imply that the channels were scoured by strong flows. Some casts with two "waterways" on the sole (Plate II, fig. 5) are indicative of channel branching. Infilling of the channels proceeded under conditions of rippling migration in a unidirectional flow as evident from occurrence of multistoried cross-lamination. Ripple marks on boundary planes between cross-bedded packages often have their crests flattened that is an argument in favor of episodic breaks in terrigenous influx and cessation of flow activity. Spill-over ripple marks suggest oscillatory pattern in deceleration of wave activity and open shallow-water settings. Measurements of cross beds (230°, n = 15), as well as trends of gutter casts (240°, n = 21) and channels (250°, n = 22) indicate that the source of terrigenous material was situated in the northeast.

Interbedded sandstone, siltstone, and mudstone characterizing the rest of the formation accumulated in a flat shallow-water plain favorable for progressive sorting of sedimentary material by waves, as evident from widespread symmetrical ripple marks. Sandstone packages in this sequence exhibit planar lamination and current-formed parting lineation, and suggest acceleration and discharge of turbid flows in shoal settings. Judging from morphology of cross sets, the filling of distributary channel developed in the course of migration of underwater dunes under influence of a long-term unidirectional flow. When cross-bedded intervals grade upward into sandstone beds with planar lamination (Plate II, fig. 6), the transition can be explained by the flow acceleration and shoaling. The general red coloration of deposits and siderite concretions are indicative of brackish-water environments (Pirrus, 1992). Measurements of cross beds (240°, n = 5) and trends of current-formed parting lineation (215°, n = 8) suggest the flow direction from the northeast.

To summarize, the estuarine and mouth bar areas of sedimentation in the Zimnie Gory time evolved into marine prodelta settings of the Erga epochs. The cyclothems resemble cyclic stratification of deltaic, schlieren, and molasse formations and suggest the analogous sedimentological mechanism that was controlled by the prodelta itself. The further evolution was associated with gradual transformation of marine prodelta settings into those of a flat freshwater plain extending away from the shore. Freshwaters flooded the plain via the system of multiple distributary channels and interacted with seawater of the offshore zone. Depositional settings described above characterize altogether a delta plain, and sedimentary succession of the Erga Formation recorded the history of that delta plain progradation from the northeast (Fig. 4). According to characteristics presented above, the prodelta deposits at the base of the Erga Formation correspond to the highest sea-level stand in the Zimnie Gory-Erga succession (Fig. 4).

Genetic Interpretation

Compositional and structural features of the deposits described above has lead to the conclusion that the studied Vendian Complex is accumulated mostly in settings of the delta-front platforms, whereas the delta tracts proper were situated northeastward of the studied outcrops. This is evident from the unimodal distribution of paleocurrent directions that is established based on measurements of cross bedding and trends of gutter casts, channels, and other current-formed structural elements. In general, it was a basin with an active hydrodynamics as evident from sedimentary material transport throughout the southeastern White Sea area and wide distribution of cross- and ripple cross-lamination in the well-sorted deposits. The both types of lamination characterize a complex circulation system with energetic transverse and along-shore currents in the open, relatively shallow offshore zone. The genetic analysis of diverse lithologic types shows that clastic material was transported into basin by long-lasted unidirectional flows.

As is established, flooding events in large deltas are associated with development of turbid flows, many of which are quasistationary, running over dozens of hours (Normark and Piper, 1991; Myrow and Southard, 1996; Mulder *et al.*, 1998). Concentration of material suspended in these flows is sufficient for sediment traction over the basin floor. The turbid flow models and processes observable in deltas show that inertia driven plumes are able to scour the basin floor and to form deposits of the turbidite type (Mulder *et al.*, 1998). The driving force of these hyperpycnal plumes (Normark and Piper, 1991) is inertia of river flows running into basin during flooding events in response to seasonal atmospheric precipitations, or after a rapid thawing of show and glaciers (Syvitski and Farrow, 1983; Maizels, 1989; Russel and Knudsen, 1999). Based on these data, I would like to compare the turbid flows, which transported sandy material to Vendian depositional settings in the southeastern White Sea area, with the hyperpycnal currents associated with flood events.

Thus, the fluviomarine deposits of the offshore sandy shoal near river mouth can be regarded as submarine or basinal alluvium. Their bedding style is likely to be of the flood origin, and the deposits proper can be regarded as inundites. Even, gently sloping seafloor relief in the offshore zone and high inertia of flows facilitated progradation of flood deposits far away from the shore. Along the delta front and strandplain flanks, there were prodelta settings, where sediments accumu-

lated under influence of interacting river stream and rough sea. Prodelta represented a submarine platform suitable for delta progradation, indicative of which are gutter casts of variable size formed under influence of hyperpycnal inertia-driven plumes related in origin to flood events, and the cast fills consequently represent inundites. Absence of terrestrial vegetation conditioned the distinct event stratification of Vendian Complex in the southeastern White Sea area, and sedimentation of that time resembled sedimentation in offshore zones of recent sea basins situated in arid and Arctic regions (Syvitski and Farrow, 1983; Maizels, 1989; Russel and Knudsen, 1999).

Characteristic of Vendian deposits is the shagreen under surface of siltstone and sandstone beds ranging in thickness from 1–2 mm to 0.5 m (Plate II, figs. 7 and 8). Being independent of the beds' genesis, the shagreen under surface is however untypical of gutter casts and sandstone beds with cross, planar, or hummocky stratification, which have been deposited in settings with most active hydrodynamics. On the other hand, soles of this kind are characteristic of beds exemplifying the prodelta settings. Small load casts (Plate II, fig. 7) and tensile-stress folding (Plate II, fig. 8) can be viewed on the shagreen soles incrusted sometimes with fine pyrite grains and distinguishable pyritized remains of algal filaments (Plate II, fig. 7). Being entangled in meshwork, the remains cover considerable areas of the soles and appear to be buried in situ. According to these observations, the shagreen soles may represent impressions of microbial films (Seilacher *et al.*, 1985; Gerdes *et al.*, 2000; Gehling, 2000). These films presumably protected bottom sediments from scouring, as the shagreen soles are lacking erosion marks.

CONCLUSION

In general, the Vendian Complex of the southeastern White Sea area corresponds to the regressive sedimentary succession deposited in the offshore zone near river mouth, when the deltaic distributary system gradually prograded from the northeast, away from the Kanin-Timan fold-and-thrust belt. The incipient formation stage of the succession was associated with a quick flood and origination of strandplain settings. Except for a thin gravelstone cover at the base of the Lyamtsa Formation, there are no other records indicative of that flood. Sediments characteristic of ingression events have not been established near the succession base. Accordingly, the lower Vendian boundary is of transgressive origin (Fig. 4).

The analyzed structure and depositional environments of Vendian sediments show that the main boundary in the succession corresponds to the surface between the Verkhovka and Zimnie Gory formations and has been associated with scour and fill of a submarine valley. This boundary separates the Lyamtsa-Verkhovka and Zimnie Gory-Erga subcomplexes, each representing a tract of sedimentary system (Fig. 4).

According to the facies architecture, the Lyamtsa-Verkhovka Subcomplex is divisible into three units that correspond principal phases of sedimentation. The thick basal progradational unit of sandy to clayey deposits records transition of strandplain environments (Lyamtsa Formation) into the mouth bar settings (lower part of the lower Verkhovka mesocycle). The overlying clayey-sandy unit of fluviomarine deposits (upper part of the aforementioned mesocycle), therefore, marks the phase of the low sea-level stand. The upper unit of the subcomplex is composed of sandy-clayey deposits of the prodelta settings (upper mesocycle of the Verkhovka Formation). The whole subcomplex can be regarded as the regressive-transgressive tract of the mouth bar sedimentary system (Fig. 4).

The sedimentary succession of the Zimnie Gory-Erga Subcomplex also includes three units. The basal sandy-clayey one is represented by estuarine and mouth-bar sediments (Zimnie Gory Formation), the fill of a submarine valley. Sandy-clayey cyclothems of the overlying unit accumulated in prodelta settings during the phase of a high sea-level stand (lower mesocycle of the Erga Formation). Finally, the thick uppermost unit is composed of clayey to sandy sediments of the delta plain (middle and upper mesocycles of the Erga Formation). As a whole, the subcomplex can be attributed to the transgressive-regressive tract of the delta plain sedimentary system (Fig. 4).

The boundary between the Verkhovka and Zimnie Gory formations is lacking evidence for subaerial exposure. It is hardly admissible, therefore, that the submarine valley was scoured in response to a sea-level drop by the end of the Verkhovka time. Immediately below the erosion surface, there are distal prodelta deposits of the upper Verkhovka mesocycle, and this is as well inconsistent with the area emergence at that time. There are two possible explanations for scouring at the base of the Zimnie Gory Formation: it could be a result of the nearshore slope erosion during a quick transgression, or a consequence of the offshore marine erosion in response to decline of terrestrial influx (Swift *et al.*, 1987; Galloway, 1989). The boundary separates two sedimentary system tracts and likely can be defined as the sequence boundary (Fig. 4).

In addition to this boundary, there are two transgressive boundaries distinguishable in the Vendian succession (Fig. 4). The lower one is conformable, and it originated in response to the northeastward shift of depositional system. This boundary separates the siltstone-mudstone sequence with sandstone interbeds of wave-agitation zone (Lyamtsa Formation) from overlying thin-bedded mudstone (Verkhovka Formation) deposited in relatively quiet waters (Fig. 4). Deposits of the Verkhovka Formation have accumulated in the course of gradual progradation of the mouth bar under hydrodynamic conditions, which were more active than hydrodynamics of the strandplain progradation during the accumulation epoch of the Lyamtsa Formation.

Moreover, the early diagenetic concretions widespread in lower beds of the Verkhovka Formation represent a characteristic feature of the important marker horizon traceable along the northeastern margin of the East European platform to the western flank of the Urals (Kirsanov, 1968). The concretions marking a regional event were probably formed in sediments at the time of low sea-level stand and maximum progradation of the mouth bar. Thus, the boundary between Lyamtsa and Verkhovka formations divides two depositional systems (those of strandplain and mouth bar) within the sedimentary system tract. As corresponding to the beginning of the new sedimentary megacycle, it can be regarded as the subsequence boundary (Fig. 4).

The upper transgressive boundary is between the Zimnie Gory and Erga Formations. The regressive mesocycles underneath change into transgressive ones above this boundary suggesting a sharp break in the sedimentary system evolution (Fig. 4). The boundary separates the estuarine and mouth bar sediments (the submarine valley fill) from overlying prodelta sediments deposited under conditions of the high sea-level stand. Deep (1 to 5 m) incisions in underlying deposits, which are characteristic of this boundary, are filled with fine-grained sandstone enclosing pebbles. These pebbles are in imbricated position within fine-grained sandstone matrix. Their drag marks and high frequency of gutter casts at the interface are indicative of the recurrent intense wave agitation and condensed sedimentation. It is likely that transgression exhumed the fluvial pebbly accumulations and redistributed pebbles in a form of lenticular intercalations. Thus, the lower boundary of the Erga Formation corresponds to the maximum flood event and divides two depositional systems (those of mouth bar and prodelta) within the sedimentary system tract. As marking simultaneously the commencement of the new sedimentary megacycle, it can be viewed as the subsequence boundary (Fig. 4).

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The Azeri, Lasnamagi, and Uhaku Ordovician Horizons of the Leningrad Region and Their Trilobite and Ostracode Faunas

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Abstract—Data on lithology, trilobites, and ostracodes from the Azeri, Lasnamagi, and Uhaku horizons of the Ladoga structural–facies zone are discussed. Biostratigraphic subdivision based on asaphids (trilobites) has been conducted, and the Azeri ostracode assemblage including over 20 species has been distinguished. The new ostracode species *Steusloffia volkhovi* sp. nov. is described and most characteristic trilobite and ostracode taxa are illustrated.

Key words: Azeri, Lasnamagi, and Uhaku horizons; Simankovo, Duboviki, Porogi, Valim, and Velesy formations; stratigraphy, trilobites, ostracodes.

This paper continuing our previous publication (Ivantsov and Mel'nikova, 1998) is aimed to complete description of formation stratotypes and biostratigraphic units of the carbonate portion of the Ordovician sequence in the Ladoga (Volkhov after Mägi, 1984) structural–facies zone.[1] In the southern Ladoga region, stratigraphic subdivision of the Azeri to Uhaku carbonate sediments, or "*Echinosphaerites* Limestone" after Schmidt, was first carried out by Raymond (1916) who distinguished the Duboviki and Revel formations in the Volkhov River section (Table). In the early 1920s, geologists from Leningrad intensely studied outcrops of the *Echinosphaerites* Limestone subsequently flooded by the Volkhov Reservoir. They collected paleontological samples predominantly from taluses. To facilitate samples' positioning, Gekker (1923) subdivided the sequence into 25 horizons or layers designated by letters from *a* to *z*. Subsequently, a stratum with ferruginous ooliths of the upper "Chechevichnyi Bed", the base of which was thought to correspond to the lower boundary of the *Echinosphaerites* Limestone in the western sections of the Leningrad region and Estonia, was distinguished, and one more layer *a1* was added to the underside of the Gekker's succession (Yanishevskii, 1931). After the mentioned works, a great collection of organic remains and important geological information accumulated, but the results have not been published. Exceptions are a paper by Gekker devoted to one cystoid group and the aforementioned brief characterization of *Echinosphaerites* Limestone by Yanishevskii. Balashova and Balashov (1961) who studied archives of A.F. Lesnikova found out that she distinguished trilobite zones and brachiopod beds in formations of the Volkhov section: the *A. eichwaldi, A. cornutus*, and *A. kowalewskii* zones in the Volkhovstroi (Duboviki) Formation; the *A. ornatus* Zone in the Porogi Formation; the *A. devexus* var. *applanata* Zone in the Valim Formation; and the *Porambonites janischevskii* and *Christiania oblonga* beds corresponding in the range to the Velesy Formation.

The *k* and *q* beds of the Gekker's scheme are best identifiable, as they are composed of dense massive cavernous limestones forming pronounced ledges in the outcrops. The base of Bed *k* corresponds now to the lower boundary of the Lasnamagi Horizon, and the top of Bed *q* to that of the Uhaku Horizon (Ivantsov, 1997). Bed *a1* distinguished by Yanishevskii approximately corresponds to the lower (first) and middle (second) members of the Simankovo Formation, both referred to the Kunda Horizon (Ivantsov, 1993). Beds *a* to *h* and *p* are most rich in trilobite remains. In the upper part of Bed *t*, there are large lenses up to 20–30 m long, composed of *Xenasaphus devexus* carapaces overlying each other in several strata. The sloughed trilobite carapaces in these accumulations are of identical size, likely belonging to male and female forms.

The top of the Ordovician sequence is strongly eroded in the Volkhov region. The overlying Upper Devonian sandstones cover different layers in separate outcrops. The Devonian and, partly, Ordovician sediments were eroded, in turn, during the Quaternary, and some outcrops known in the early 20th century are nowadays covered by waters of the hydroelectric sta-

[1] We take an opportunity to apologize for errors in the published work (Ivantsov and Mel'nikova, 1998), where one should read *Trypanites* instead of *Bergaueria* throughout the text, except for the figure captions, and the glauconite-enriched levels in Fig. 1 for the Babino section are erroneously referred to as oolitic beds.

Table 1. Correlation of subdivision schemes suggested based on trilobites from the Azeri-Uhaku sequence in the Ladoga structural-facies zone

Lamanskii, 1901, 1905	Yanishevskii, 1931	Lesnikova (from Balashova, Balashov, 1961)	Balashova and Balashov, 1959, 1961			*Resheniya..., 1987*		Ivantsov (present paper)		
Horizon (zone)	Horizon	Zone	Horizon	Subhorizon	Zone	Subzone	Horizon	Beds	Horizon	Zone (beds)	Other trilobites
				Velesy			Uhaku	Chasmops odini, Illaenus schmidti, I. intermedius	Uhaku		
	A. devexus	*A. devexus*	Duboviki	Valim	*A. devexus*		Uhaku	*X. devexus*	Uhaku	*Xenasaphus*	*X. devexus* *X. mjannili*
	Asaphus sp.	*A. ornatus*		Porogi	*A. ornatus*		Lasnamagi	*I. schroeteri*, *A. ornatus*	Lasnamagi	*A. bottnicus*	
	A. kowalewskii, *A. latus*, *A. laevissimus* var. *laticauda*	*A. kowalewskii*		Volkhovstroi	*A. kowalewskii*	*A. kowalewskii*	Azeri	*A. kowalewskii*, *A. cornutus*, *A. platyurus*	Azeri	*A. kowalewskii*	
	A. cornutus					*A. kowalewskii*, *A. cornutus*				*A. intermedius*	
Asaphus platyurus, *A. kowalewskii*		*A. cornutus*			*A. cornutus*					*A. punctatus*	
	A. eichwaldi, *A. eichwaldi* aff., *A. laevissimus*, *A. cornutus*, *A. cornutus* var.					*A. kowalewskii*, *A. cornutus*, *A. eichwaldi*				*A. kotlukovi*	
		A. eichwaldi			*A. eichwaldi*	*A. cornutus*, *A. eichwaldi*				*A. heckeri*	*S. laticaudatus* *S. latisegmentatus*
A. eichwaldi, *Ptychopyge globifrons*						*A. eichwaldi*, *A. minor*	Kunda	*A. sulevi*, *P. globifrons*, *Megistaspis gigas*, *M. obtusicauda*	Kunda	*A. laevissimus*	
										A. ingrianus	
										A. sulevi	
										A. pachyophthalmus	
										A. minor	
			Obukhovo		*A. major*					*A. striatus*	*P. globifrons* *A. knyrkoi*
A. raniceps					*A. raniceps*			*A. "raniceps"*, *M. lawrowi*		*A. raniceps*	*Pliomera fischeri*
A. expansus, *A. lamanskii*					*A. expansus*, *A. lamanskii*			*A. expansus*		*A. expansus*	*A. lamanskii* *I. "esmarki"*

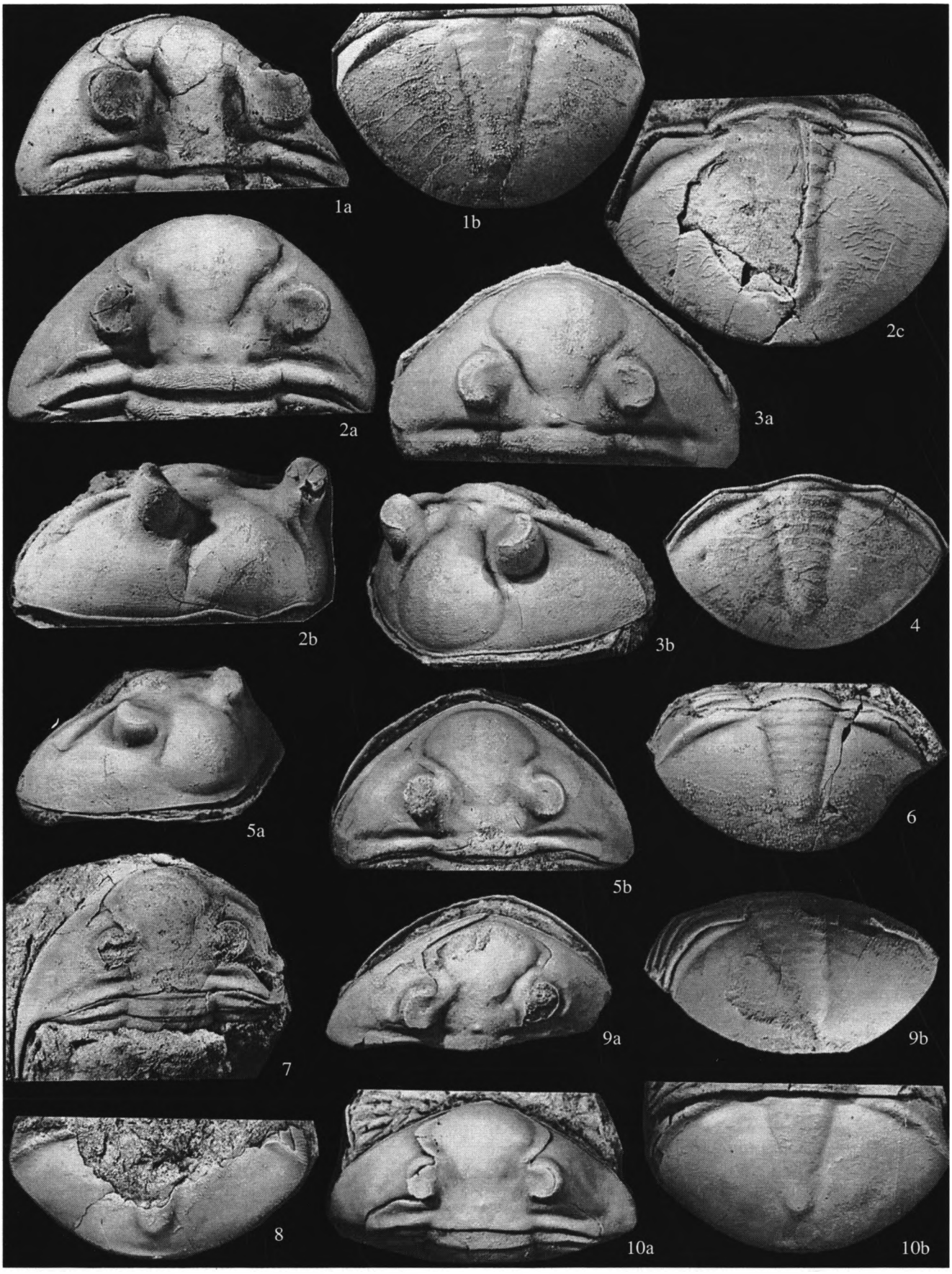
tion reservoir. As a result, beds *w*, *x*, *y*, and *z* are inaccessible for observations. The Ordovician part of the section is terminated by Bed *v* exposed in the Bor Village nearby the railway bridge.

Revising the Lesnikova's scheme, Balashova and Balashov (1961) distinguished in the upper part of the carbonate sequence the Duboviki Horizon including the Volkhovstroi (with the *A. eichwaldi*, *A. cornutus*, and *A. kowalewskii* zones), Porogi (with the *A. ornatus* Zone), and Valim (with the *A. devexus*, *Caryocystites*, and *Ancistroceras* zones) subhorizons. Following Gekker, they attributed an upper half of the Velesy Formation in nomenclature by Lesnikova (under name of the Velesy Subhorizon) to the Kuker Horizon.

In 1984, when preparing materials to the Interdepartmental Stratigraphic Conference on the Ordovician and Silurian, Popov and Männil revised stratigraphy of the carbonate Ordovician sediments of the region (*Resheniya...*, 1987). The stratigraphic scheme they suggested for the Leningrad distribution area of the Baltic-Ladoga glint included the Azeri, Lasnamagi, and Uhaku horizons distinguished in Estonia, and the Obukhovo Formation equal in range to the Obukhovo and lower half of the Duboviki horizons in terminology of Balashova and Balashov. The trilobite zones they distinguished following Lesnikova were excluded from the scheme.

In the 1990s, one of us studied stratigraphy and distribution of asaphid trilobites in the Ordovician carbonates of the region (Ivantsov, 1993, 1996, 1997a, 1997b; Ivantsov and Mel'nikov, 1993; Ivantsov and Mel'nikova, 1998). The major result was zonal subdivision of the upper Volkhov-Azeri portion of the sequence based on the distribution of two trilobite lineages of the genus *Asaphus*.

The lower boundary of the Azeri Horizon used to be placed at the appearance level of the *Asaphus platyurus* (Männil, 1963a, 1963b, 1966, 1990; *Resheniya...*, 1987). However, this form is unsuitable for detailed stratigraphy, as it actually represents a group of species whose ranges span a greater part of and even the whole Azeri Horizon. This group was distinguished as the new genus *Platyasaphus* (Ivantsov, 2000). As the genus includes the type species of the subgenus *Asaphus* (*Subasaphus*) Balaschova (Balashova, 1976), the name *Platyasaphus* Ivantsov should be considered as the junior synonym of *Subasaphus* Balaschova. Subasaphids from the Ladoga region considerably differ from Swedish forms, at least from the neotype *A. platyurus* (Jaanusson, 1953), and two groups likely existed at differ-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ent times. *S. latisegmentatus* was encountered at the base of the Azeri Horizon and *S. spinifer* in its middle part. Judging from specimens found by private collectors, subasaphids with buccal spines (=*A. platyurus* s.l.) occur higher upward but without passing into the Lasnamagi Horizon.

The new species *Asaphus heckeri* (Plate I, fig. 9) was recently described from the lower part of the Azeri Horizon (Ivantsov, 2000). According to morphological features, it is a form intermediate between *A. laevissimus* and *A. kotlukovi*, representing a connecting link between the *A. sulevi–A. ingrianus–A. laevissimus* and *A. kotlukovi–A. punctatus–A. intermedius–A. kowalewskii* lineages. All species of both lineages are zonal forms (Ivantsov, 1996, 1997a, 1997b). Their stratigraphic range corresponds in total to the upper third of the Kunda Horizon coupled with the greater portion of Azeri Horizon, though with a gap at the base of the latter. In the previously proposed scheme, this gap was occupied by *S. latisegmentatus* whose origin is unclear, and whose range may be wider than that suggested by records from the Baltic-Ladoga glint. Thus, we consider now *A. heckeri* as the zonal species of the lowermost Azeri Horizon. This form was encountered as yet only in the sections nearby the Ladoga Lake southern shore and at two points in the Baltic part of the glint. The upper boundary of the zone is marked by appearance of *A. kotlukovi*, the assumed nearest descendant of *A. heckeri.*

Stratigraphic ranges of *A. ingrianus, A. kotlukovi, A. punctatus*, and *A. intermedius* are extremely narrow and suitable for a high-resolution correlation. However, a gradual morphological transition between the related forms *A. sulevi* and *A. ingrianus, A. kotlukovi* and *A. punctatus*, and *A. intermedius* and *A. kowalewskii* is an obstacle for identification of biostratigraphic boundaries. Therefore, ranges of these species are combined into "summary" zones in the proposed biostratigraphic scheme.

The *A. sulevi–A. kowalewskii* lineage appeared in the late Kunda time and evolved to the latest Azeri time. The main tendency of morphological changes within it was the gradual increase of eyes elevation above carapace that terminated by formation of extremely long ommatophores characteristic of *A. kowalewskii.* Despite the obvious distinction between species spaced in time, transitions between neighboring taxa, as they are recorded by fossil specimens, were nearly always gradual (specimens illustrated in Plate I, figs. 2–6, 9, and Plate II, fig. 5, are most typical, and all transitional forms are omitted). The preceding *A. lepidurus–A. pachyophthalmus* lineage was also lengthy, similarly characterized by gradual morphological changes. The sole exception is a considerably sharp transition from *A. striatus striatus* to *A. pachyophthalmus*, which is most likely indicative of a significant stratigraphic and, evidently, time interval separating them (Ivantsov, 1997a; Ivantsov and Mel'nikova, 1998). The drastic replacement of one continuous asaphid lineage by the other one that took place in the last third of the Kunda time is more surprising. An insignificant interval between findings of last *A. pachyophthalmus* and first *A. sulevi* (Fig. 3) in no way corresponds to the degree of distinction between the species (Plate I, figs. 1 and 2). According to its morphological features, *A. sulevi,* the ancestor of new lineage, cannot be directly derived from any member of the preceding lineage and from other known *Asaphus* forms. The sharp morphological changes, which took place in the interval between the *A. pachyophthalmus* and *A. sulevi* existence periods, are more significant than transformations across the Kunda-Azeri boundary assumed at present. *Pliomera fischeri* characteristic of the Kunda Horizon became extinct exactly at this time, and typical Kunda forms of the *Illaenus "esmarki"* group died out slightly earlier. It is possible that faunal assemblages in the Baltic basin experienced their transformation precisely at that time to get the appearance of Azeri assemblages.

Ostracodes also illustrate evolutionary changes, which preceded environmental transformations. To

Plate I. Trilobites from the Kunda and Azeri horizons.

(1) *Asaphus pachyophthalmus* F. Schmidt, 1901, specimen no. 4330/36 (×1.5), cephalon (a) and pygidium (b); Outcrop 3b, left bank of the Volkhov River, Volkhov town, between the railway bridge and the VHES dam; Kunda Horizon, 1.9 m below the top, Simankovo Formation, Member 1, the *minor-pachyophthalmus* Zone. (2) *Asaphus sulevi* Jaanusson, 1953, specimen no. 4330/46 (×1.5), cephalon (a, b) and pygidium (c); Outcrop 3a, left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, about 0.5 km downstream of outcrop 3b; Kunda Horizon, 1.1 m below the top, Simankovo Formation, Member 1, the *sulevi-ingrianus* Zone. (3, 4) *Asaphus ingrianus* Jaanusson, 1953: (3a, 3b) specimen no. 4330/81, cephalon(×2), left bank of the Volkhov River near the northern margin of the Plekhanovo Village, Kunda Horizon, Simankovo Formation, Member 1, the *sulevi-ingrianus* Zone, talus; (4) specimen no. 4330/82, pygidium (×1.2); Outcrop 3b, Kunda Horizon, 0.55 m below the top, Simankovo Formation, Member 1, the *sulevi-ingrianus* Zone. (5, 6) *Asaphus laevissimus* F. Schmidt, 1901: (5a, 5b) specimen no. 4330/47, cephalon (×1.5), Outcrop 3b, Kunda Horizon, 0.2 m below the top, Simankovo Formation, Member 2, the *laevissimus* Zone; (6) specimen no. 4330/85, pygidium (×2), ditto, the same age, 0.15 m below the top of the Kunda Horizon. (7, 8) *Subasaphus latisegmentatus* (Nieszkowskii, 1857): (7) specimen no. 4330/92, cephalon of a form with elongated carapace (×1.1), Outcrop 3b, Azeri Horizon, Simankovo Formation, Member 3, talus, the *heckeri* Zone, *latisegmentatus* Subzone; (8) specimen no. 4330/91, pygidium of a form with shortened carapace (×2), Volkhov River, collection of Balashova. (9) *Asaphus heckeri* Ivantsov, 2000, specimen no. 4330/48 (×1.5), cephalon (a) and pygidium (b); Outcrop 3b; Azeri Horizon, Simankovo Formation, Member 3, talus, the *heckeri* Zone. (10) *Subasaphus laticaudatus* (F. Schmidt, 1901), specimen no. 4330/44 (×1), cephalon (a) and pygidium (b); Outcrop 3a; Azeri Horizon, 1.1 m above the base, Simankovo Formation, Member 3, the *heckeri* Zone, *laticaudatus* Subzone.

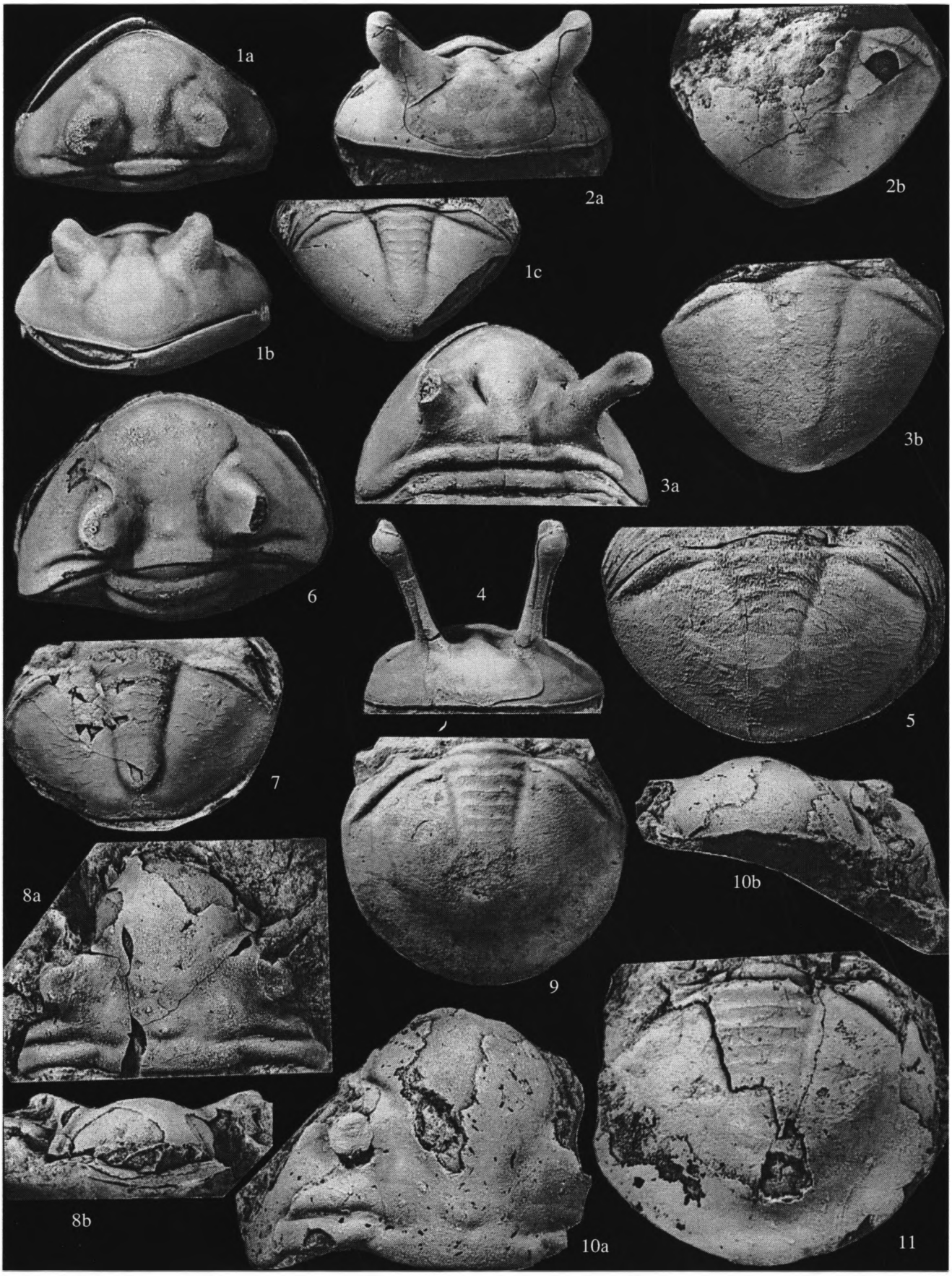
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Plate II. Trilobites from the Azeri, Uhaku, and Lasnamagi horizons.

(1) *Asaphus kotlukovi* Balashova, 1953, specimen no. 4330/98, a form with highly elevated eyes (×1.5), cephalon (a, b) and pygidium (c); Outcrop 3b; Azeri Horizon, Simankovo Formation, Member 3, the *kotlukovi-punctatus* Zone, talus, collection of Lesnikova. (2) *Asaphus punctatus* Lesnikova, 1949, specimen no. 4330/58, form with long ommatophores (×1), cephalon (a) and pygidium (b); Outcrop 3b; Azeri Horizon, Duboviki Formation, the *kotlukovi-punctatus* Zone, talus. (3) *Asaphus intermedius* Balashova, 1953, specimen no. 4330/63, (×1.5), cephalon (a) and pygidium (b); Outcrop 4, right bank of the Volkhov River, Volkhov town, 100 m downstream of the VHES lock; Azeri Horizon, Duboviki Formation, 4.4 m below the top, the *intermedius-kowalewskii* Beds. (4, 5) *Asaphus kowalewskii* Lawrow, 1856 (×1.5): (4) specimen no. 4330/61, cephalon, Outcrop 3a, Azeri Horizon, Duboviki Formation, about 2 m below the top, the *intermedius-kowalewskii* Beds; (5) specimen no. 4330/62, pygidium of the form with a shortened carapace, Outcrop 3b, Azeri Horizon, Duboviki Formation, about 2.35 m below the top, the *intermedius-kowalewskii* Beds. (6, 7) *Asaphus bottnicus* Jaanusson, 1953: (6) specimen no. 4330/125, cephalon(×2), Outcrop 5b, left bank of the Volkhov River, about 1.2 km upstream of the VHES dam, Lasnamagi Horizon, Porogi Formation, 5.3 m above the base, the *bottnicus* Beds; (7) specimen no. 4330/143, pygidium (×1.5), ditto, 5.15 m above the base of the Porogi Formation. (8, 9) *Xenasaphus (Xenasaphus) mjannili*, Ivantsov, 2000: (8) specimen no. 4330/136, cephalon (×1), dorsal view (a), front view (b); (9) specimen no. 4330/138, pygidium (×0.75), Outcrop 6b, left bank of the Volkhov River near the Valim Village, Uhaku Horizon, Valim Formation, 2–2.1 m above the base of the formation, the *Xenasaphus* Beds. (10, 11) *Xenasaphus (Xenasaphus) devexus* (Eichwald, 1840): (10) specimen no. 4330/134, cephalon (×1), dorsal view (a), front view (b); (11) specimen no. 4330/135, pygidium (×0.75), Outcrop 6b, Uhaku Horizon, Valim Formation, 3.9–4.0 m above the base of the formation, the *Xenasaphus* Beds.

trace changes in ostracode fauna across the Kunda-Azeri boundary in the eastern Leningrad region, we collected samples from the upper part of the Kunda and throughout the Azeri horizons. Ostracodes from the Azeri sediments are almost invisible, and we discovered scarce, though taxonomically diverse, ostracode remains only after alloying the rocks with sodium hyposulphite. It should be noted that the Kunda ostracodes of the Baltic paleobasin are well known (Netskaya, 1953; Sarv, 1959; Sidaravichene, 1992; Schallreuter, 1993; Ivantsov and Mel'nikova, 1998), whereas information on the Azeri forms is more limited. For instance, Sarv (1959, p. 192) reported that the ostracode assemblage from the Azeri Horizon of Estonia is very impoverished, consisting of four species, and Sidaravichene (1992, p. 205, Table 2) reported about five taxa from this interval in Lithuania. In the westernmost Leningrad region, namely, in the distribution area of combustible schists deposit, Sarv found six species from the Azeri sediments recovered by Well 1146 (Sarv *et al.*, 1985).

The sections described below are variously characterized by ostracodes, distribution of which is illustrated in Figs. 2 to 4 and 7.[2] The Azeri ostracode assemblage is represented by the following species: *Gryphiswaldensis* sp. A (Plate III hereafter, fig. 20), *Euprimites effusus* Jaanusson, 1957 (fig. 17), *Uhakiella cicatriosa* Sarv, 1959 (fig. 15), *Eobromidella zvankensis* Meln., 1998 (fig. 7), *Tallinnella marchica* (Krause, 1889) (fig. 13), *Glossomorphites* aff. *acutus* (Hessland, 1949) (fig. 4), *Easchmidtella ivantsovi* Meln., 1998 (fig. 19), *Steusloffia volkhovi* sp. nov. (fig. 5), *Tetrada (Perplana) perplana* (Neckaja, 1953) (fig. 23), *Piretella oepiki* Thorslund, 1940 (fig. 18), *Rakverella (Pectidolon) perornata* (Opik, 1937) (fig. 22), *Collibolbina plana* (Hessland, 1949) (fig. 21), *Sigmobolbina* sp. (fig. 12), *Longiscula arcuaris* Neckaja, 1958 (fig. 10), *L. affluensis* Olempska, 1994 (fig. 8), *L. parrectis* Neckaja, 1958 (fig. 9), *L.* sp. (fig. 11), *Leperditella semen* Sarv, 1959 (fig. 1), *Baltonotella ledai* Sidaravichiene, 1975 (fig. 16), *Ahlintella pseudofabuliformis* Sidaravichiene, 1994 (fig. 3), *Steusloffina* aff. *cuneata* (Steusloff, 1895), *Bolbihithis* sp. (fig. 2), *Primitiella molli* (Bonnema, 1909) (fig. 14), and *Bollia?* sp. (fig. 6).

Ostracodes studied in the eastern Leningrad region considerably extended the list of taxa characterizing the Azeri assemblage. The analyzed stratigraphic distribution of ostracodes indicates that certain species, previously recognized as characteristic of the younger Ordovician horizons in the Eastern Baltic structural–facies zone, appear already in the Azeri strata of the sections under discussion. For instance, *Rakverella (Pectidolon) perornata* is known from the Uhaku and Kukruze horizons of Estonia and from the Idavere and Iykhvi horizons of Lithuania. *Tetrada (Perplana) perplana* characterizes the Kukruze Horizon of the stratotype area in Estonia. In addition, some species of the Azeri ostracode assemblage appear in the upper part of the Kunda Horizon, below two upper trilobite zones of this subdivision (Ivantsov and Mel'nikova, 1998, p. 59, Fig. 2). Among them, there is *Piretella oepiki* identified in the Zvanka section as *Lembitsarvella* sp. known from the upper part of the *Chasmops* Beds (Uhaku) of Sweden, and from the Uhaku, Iykhvi, and Nabala horizons of Poland (Olempska, 1994). *L. affluensis* is also described from the same horizons of Poland, whereas *Longiscula parrectis* is reported from the Uhaku and Kukruze horizons of the Pskov region.

STRATIGRAPHIC SUBDIVISIONS

Azeri Horizon. The lower boundary of the horizon is placed at the appearance level of *A. heckeri* and *S. latisegmentatus.* Its upper boundary falls within the interval barren of trilobites and conventionally corresponds to the extinction level of *A. kowalewskii* (Ivantsov, 1993). The horizon includes the following subunits (upward from the base): the *heckeri, kotlukovi*, and

[2] The section on the right bank of the Koporka River, which is situated below the fortress nearby the Koporka Village, is not described here; only the sampling sites and ostracode taxa are listed (Fig. 7).

Fig. 1. Ordovician stratotype localities in the Volkhov River region: 1) Babino Quarry; (2) left bank of the Volkhov River) northern margin of the Zvanka Village; (3a, 3b) left bank of the Volkhov River within the Volkhov town area downstream of the VHES dam with exposures near the new bridge (3a) and directly near the dam (3b); (4) right bank of the Volkhov River, the Volkhov town area 50 to 150 m downstream of the dam lock; (5a, 5b) left bank of the Volkhov River 0.2 to 1 km upstream of the dam lock, the downstream (5a) and upstream (5b) parts of the outcrop; (6a, 6b) ditto near the Porogi and Valim villages, the downstream (6a) and upstream (6b) flanks of the outcrop; (7) right bank of the Volkhov River downstream of the railway bridge near the Bor Village, the interval between the bridge and nameless creek mouth (settlement areas are shaded).

punctatus zones and *intermedius-kowalewskii* beds. The horizon as a whole corresponds to the lower half of the graptolite *D. murchisoni* Zone (*Resheniya...*, 1987). In sections of the Ladoga structural–facies zone, the Azeri Horizon is represented by members 2 and 3 of the Simankovo Formation and by the Duboviki Formation sediments.

The Simankovo Formation was distinguished by present authors (Ivantsov and Mel'nikova, 1998). Its stratotype on the left bank of the Volkhov River corresponds to exposure near the new automobile bridge in the Volkhov (Outcrop 2, Figs. 1 and 2). The formation is represented by clayey limestones and marls, displaying thin to lenticular bedding and commonly intercalated with clay beds. The lower boundary of the Simankovo Formation corresponds to the appearance level of high-clayey limestones and marls in the section. The upper half of the formation includes the following members (from the base upward):

Member 2: greenish gray limestone, thick-bedded, with a slight lumpy parting and thin, sinuous, vertical burrows. About 5 cm below the top of the member, there is a recognizable surface of unconformity with a pale lilac top, which is overlain by sediments bearing scarce pebble of clayey limestone. With these specific features, the surface is traceable throughout outcrops of the Leningrad region. In the Narva River area, it grades into the lower level of two closely spaced rusty-yellow surfaces at the top of the Napas Formation. This unconformity marks the lower distribution limit of *S. latisegmentatus* and corresponds to the base of the Azeri Horizon (Orviku, 1960; Ivantsov, 1993).

Member 3: greenish to pinkish gray marl with lenticular bedding and common clay intercalations.

The Simankovo Formation is a subdivision of the Ladoga structural–facies zone. In northern Estonia, its age analogs are the Napas and the lower part of the Azeri Formation.

The Duboviki Formation was distinguished by Raymond (1916) who did not establish its holostratotype. The outcrop on the right bank of the Volkhov River nearby the Duboviki Village, immediately downstream of the dam lock of the Volkhov Hydroelectric Station (VHES) (Outcrop 4, Figs. 1, 3), is proposed to be the lectostratotype. The formation is represented by gray, pinkish to greenish gray limestone, mainly thick-bedded and as clayey as marl at some levels. The lower boundary of the formation is placed at the base of the first thick-bedded limestone bed, evidently Bed *c* (Ivantsov, 1993). The Duboviki Formation is distinguishable in all sections of the Leningrad region. Its age analog in northern Estonia is the Azeri Formation.

A. heckeri Zone spans the complete stratigraphic range of its index species (Plate I, fig. 1). The lower boundary of the zone is defined by appearance of the index species immediately above the regional surface of unconformity. The outcrop on the left bank of the Volkhov River between the VHES dam the railway bridge (Outcrop 3b, Fig. 2) is proposed as candidate for the type section. The analog of the zone in the Ladoga glint area is the upper part of Member 2 and Member 3 of the Simankovo Formation (without its uppermost part). Specimens of *A. heckeri* are also known from sediments of the Duboviki Formation in the Pavlovsk area, Leningrad region, and from the Yaagala River sites in Estonia, the Eastern Baltic structural–facies zone (Tallinn Natural Museum, specimen 91/5197).

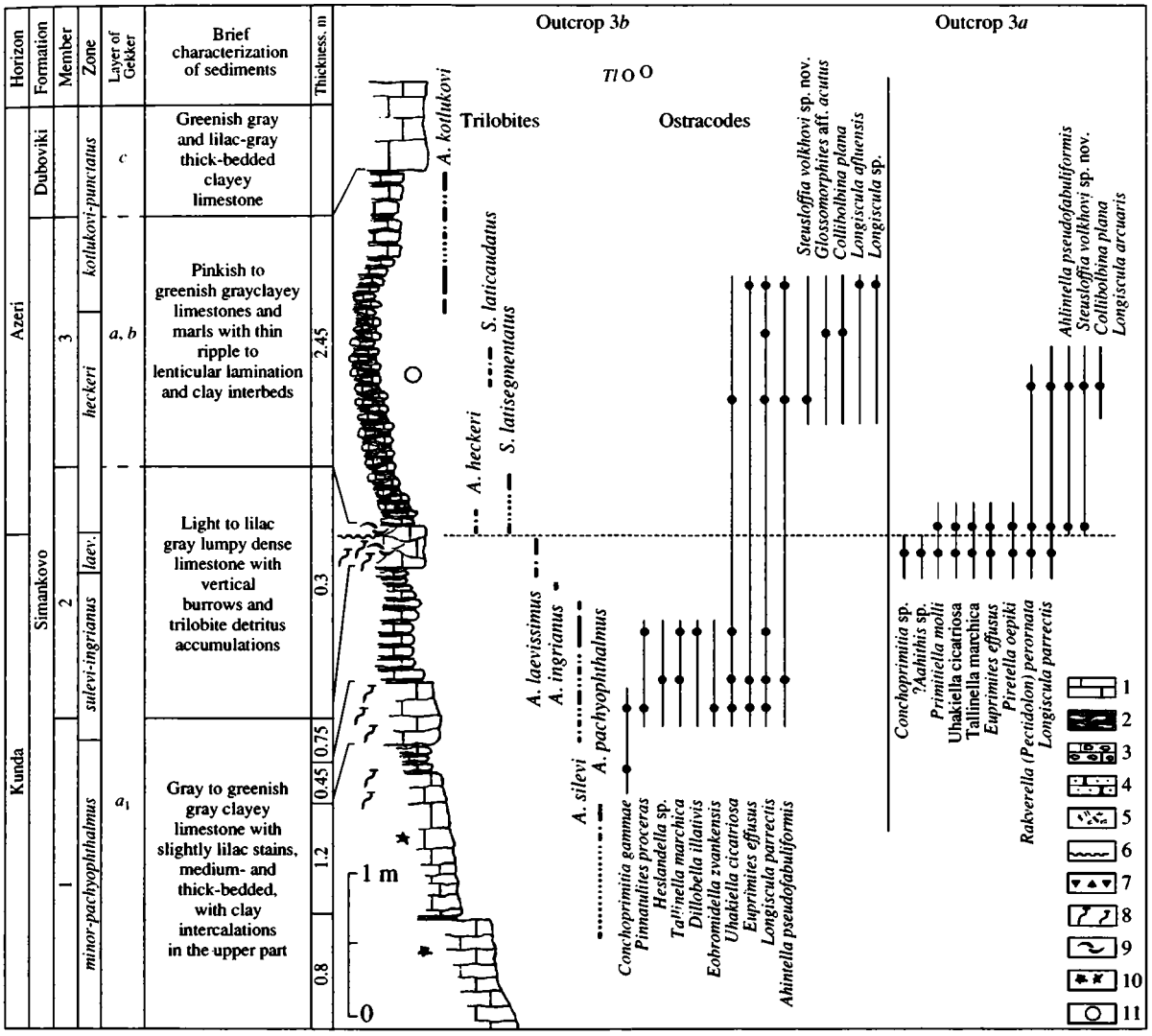

Fig. 2. Outcrop 3b, left bank of the Volkhov River immediately downstream of the VHES dam in the Volkhov town (about 0.5 km upstream of Outcrop 3a), the stratotype of the *A. sulevi, A. ingrianus, A. laevissimus, A. heckeri*, and *A. kotlukovi-punctatus* zones: (1) limestone; (2) limestone with lenticular bedding and clay interbeds; (3) limestone with leaching caverns; (4) sandstone; (5) soil; (6) unconformity surface; (7) visible glauconite grains; (8) thin and long vertical burrows; (9) accumulations of large fragments of trilobite carapaces; (10) pyrite concretions; (11) oldest *Echinosphaerites* forms; abbreviation *laev.* is for the *A. laevissimus* Zone.

A. kotlukovi–A. punctatus Zone. The zone range is equal to the combined stratigraphic ranges of *A. kotlukovi* (Plate II, fig. 1) and *A. punctatus* (Plate II, fig. 2), which are direct relatives. The outcrop on the left bank of the Volkhov River between the VHES dam and railway bridge (Outcrop 3b, Fig. 2) is proposed as a type section. The lower boundary of the zone corresponds to the appearance level of *A. kotlukovi* that is recorded in the upper part of Member 3 of the Simankovo Formation, 1.55 m above the surface of unconformity in the type section. The zone corresponds to the upper part of Member 3 of the Simankovo Formation and to the lower part of the Duboviki Formation.

A. intermedius–A. kowalewskii Beds. The unit corresponds to the combined stratigraphic range of *A. intermedius* (Plate II, fig. 3) and *A. kowalewskii* (Plate II, figs. 4 and 5). The outcrop on the right bank of the Volkhov River, 50 m downstream of the VHES lock (lectostratotype of the Duboviki Formation, Outcrop 4, Fig. 3), is proposed as a type section of the unit. The lower boundary corresponds to the appearance level of *A. intermedius*. In the type section, this species is recorded in the middle part of the Duboviki Formation, 4.45 m above the surface of unconformity. The beds span interval of the upper half of the Duboviki Formation.

Lasnamagi Horizon. It is difficult to define boundaries of the Lasnamagi and Uhaku horizons based on trilobites, because their remains are scarce and fragmentary. The authorized *A. ornatus* Beds (*Resheniya...*,

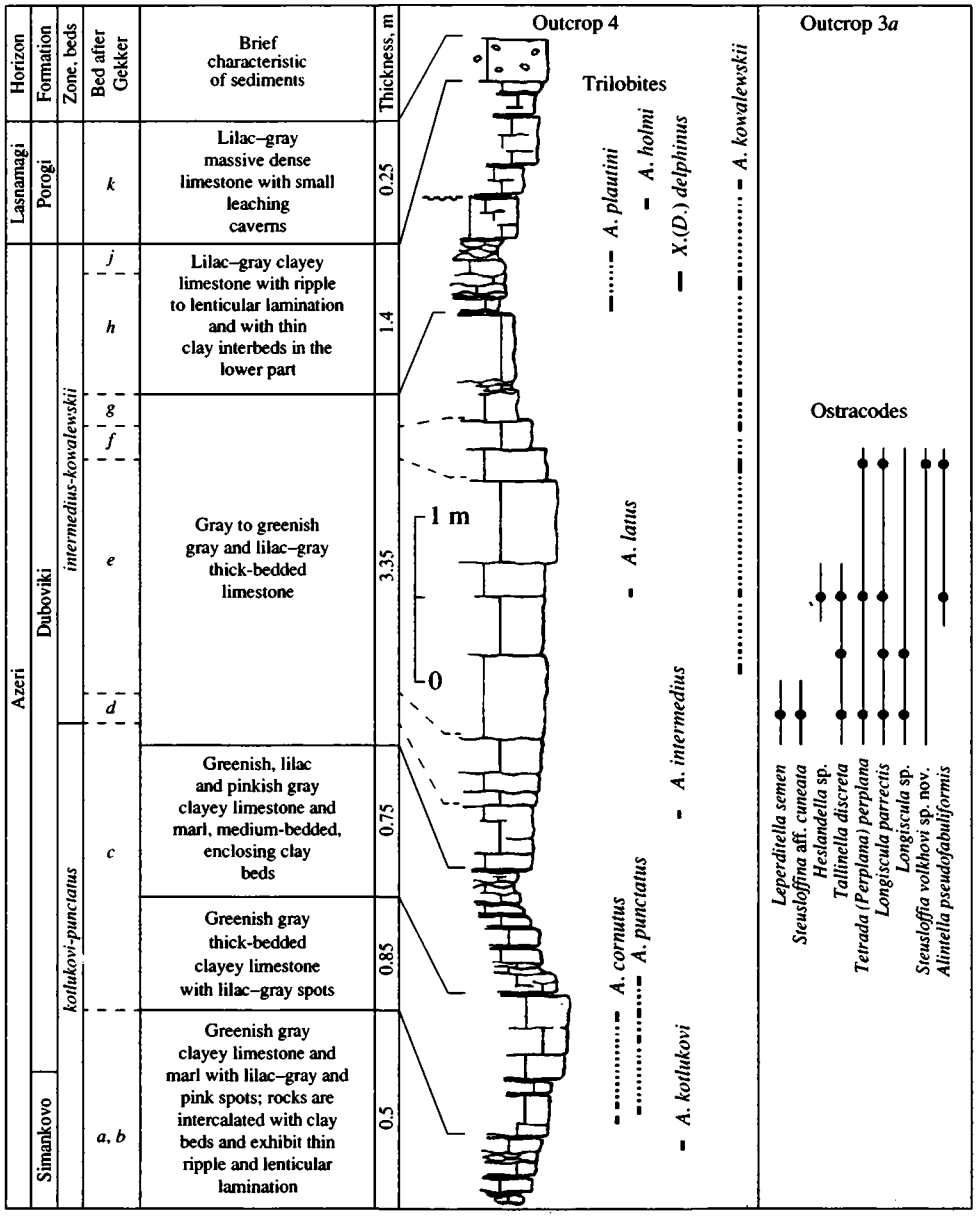

Fig. 3. Outcrop 4 on the right bank of the Volkhov River, Volkhov town area 50–150 m downstream of the VHES lock; the lectostratotype of the Duboviki Formation and stratotype of the *A. intermedius-A. kowalewskii* Beds (symbols as in Fig. 2).

1987) likely correspond only to the upper half of the Porogi Formation. Scarce trilobite remains from the lower part of the formation are almost unidentifiable. In the sections of the Ladoga structural–facies zone, the lower horizon boundary is placed conventionally at the base of the Porogi Formation, directly below which the last occurrence level of *A. kowalewskii* is recorded. In the biostratigraphic range, the horizon is equal to the upper half of the graptolite *D. murchisoni* Zone (*Resheniya...*, 1987) and includes the trilobite *A. bottnicus* Beds (Ivantsov, 1997). In the area of the Ladoga structural–facies zone, the horizon is complete analog of the Porogi Formation (*Resheniya...*, 1987).

Porogi Formation. The name was proposed by Lesnikova and published by Balashova and Balashov (1961). The holostratotype of the formation was not distinguished. The Volkhov town outcrop on the left bank of the Volkhov River, 0.2 to 1.0 km upstream of the VHES lock (Outcrop 5, Figs. 1 and 4), can be proposed as the lectostratotype. The formation is represented by lower greenish gray and upper reddish gray limestones, mainly thick-bedded, dolomitized, in places with caverns after faunal remains and pyrite concretions. Lesnikova placed the lower boundary of the formation at the base Layer *k* of the Gekker's scale that "forms a ledge in the Volkhov River sections" (Balash-

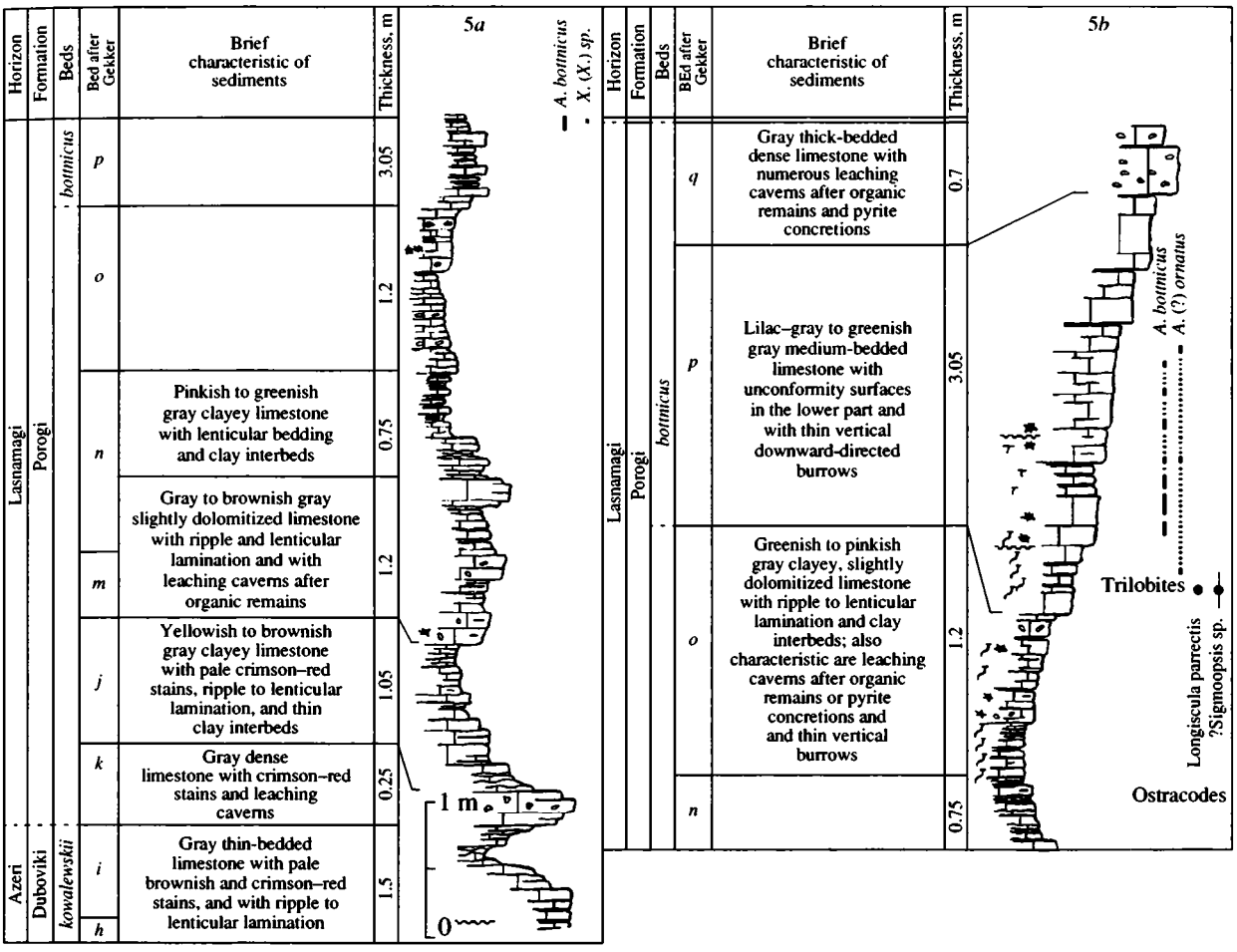

Fig. 4. Outcrop 5, left bank of the Volkhov River 0.2–1.0 km upstream of the VHES lock in the Volkhov town; lectostratotype of the Porogi Formation, stratotype of the *A. bottnicus* Beds. Symbols as in Fig. 2.

ova and Balashov, 1961). Balashova and Balashov (1961) placed the boundary one layer below, in which they encountered specimens of *A. ornatus*. These trilobite specimens are not preserved, and their decision was most likely an error, because *A. holmi* occurring along with *A. kowalewskii* in this layer could be mistaken for *A. ornatus*. *A. holmi* and *A. kowalewskii* are typical members of the Azeri trilobite fauna and are widespread in the Duboviki Formation sediments. In this paper we return to the boundary variant proposed by Lesnikova. In addition to the Volkhov River sections, sediments of the Porogi Formation are partially exposed along the Lava River and around the Putilovo Village. In northern Estonia, the age analog of the formation is the lower part of the Vyao Formation.

A. bottnicus Beds. The beds correspond to the complete stratigraphic range of *A. bottnicus* (=*A. ornatus* Pompecki sensu Schmidt 1901 (part); Plate II, figs. 6 and 7). The outcrop on the left bank of the Volkhov River 0.2–1.0 km upstream of the VHES lock (lectostratotype of the Porogi Formation, Outcrop 5, Fig. 4) is proposed as a type section. The lower boundary of the unit is defined by the appearance level of the index species; in the type section it is recorded in the middle part of the Porogi Formation, about 4.5 m above its base. Between the *A. bottnicus* and *A. intermedius–A. kowalewskii* beds, there is a 5-m-thick gap, where sediments are almost barren of trilobite remains. Few ostracodes *Longiscula parrectis* and *Sigmoopsis* sp. were found at the base of the *A. bottnicus* Beds.

Uhaku Horizon. The lower boundary of the horizon is conventionally placed at the base of the Valim Formation, as its faunal remains are as yet imperfectly studied. The horizon corresponding as a whole to the graptolite *G. teretiusculus* Zone includes the *X. (Xenasaphus)* (Ivantsov, 1997), *Chasmops odini, Illaenus intermedius*, and *I. schmidti* beds (*Resheniya...*, 1987). Within the Ladoga structural–facies zone, the horizon spans the range of the Valim and Velesy formations (*Resheniya...*, 1987).

Valim Formation. The name of the formation was proposed by Lesnikova and published by Balashova and Balashov (1961). The outcrop on the left bank of the Volkhov River nearby the Porogi and Valim villages

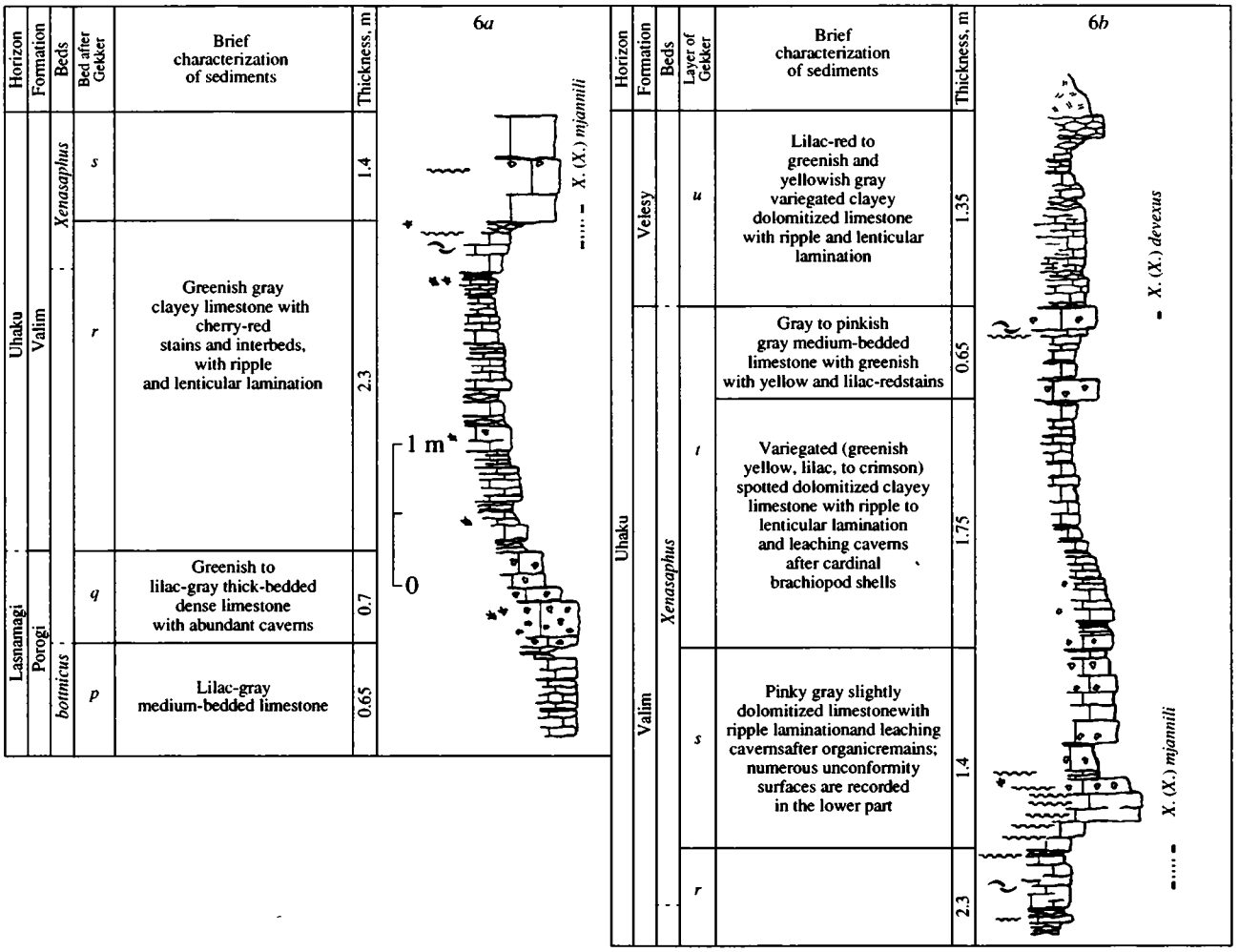

Fig. 5. Downstream (6a) and upstream (6b) parts of the outcrop on the Volkhov River left bank (sites of the Porogi and Valim villages); lectostratotype of the Valim Formation and stratotype of the *X. (Xenasaphus)* Beds (symbols as in Fig. 2).

(Outcrop 6, Figs. 1 and 5) is its stratotype. The formation is represented by clayey limestones and marls, gray to variegated, mainly reddish, dolomitized and cavernous in places. Balashova and Balashov placed the lower boundary of the formation at the top of the dense limestone bed with large caverns (Beds *q* and likely the lowermost *r*). Sediments of the formation are exposed only in the stratotype area. In northern Estonia, the age analog of the Valim Formation is the upper part of the Vyao Formation.

Velesy Formation. The name was proposed by Lesnikova and published by Balashova and Balashov (1961). The stratotype was not distinguished. Outcrops of the formation in its full range are unknown; those formerly situated nearby the Velesy Village are flooded, and Beds *w–z* were eliminated in the others by the pre-Late Devonian erosion. A relatively complete outcrop on the right bank of the Volkhov River near the Bor Village, slightly downstream of the railway bridge, below the garden strips (Outcrop 7, Figs. 1 and 6), can be proposed as a lectostratotype of the formation. Sediments of the unit are represented by clayey limestones and marls, pinkish gray and dolomitized. The lower boundary of the formation is placed at the top of Layer *t*, the upper part of which is represented by yellowish gray dolomitized limestone bearing abundant remains of *X. (X.) devexus.* In other areas, the formation is not exposed; its age analog in northern Estonia is the Kyrgekalla Formation.

The Xenasaphus Beds correspond to the stratigraphic interval characterized by the maximum abundance of trilobites *X. (Xenasaphus)* (= *Asaphus devexus* sensu Schmidt 1901 (part). The outcrop on the left bank of the Volkhov River close to the Porogi Village (lectostratotype of the Valim Formation, Outcrop 6, Fig. 5), is proposed as a type section. The lower boundary of the unit in the type section is placed at the base of the first mass accumulation of *X. (X.) mjannili* (Plate II, figs. 8 and 9) that is recorded 2.0 m above the base of the Valim Formation. The upper boundary of the unit is defined by the last mass accumulation of *X. (X.) devexus* (Plate II, figs. 10 and 11) recorded in the type

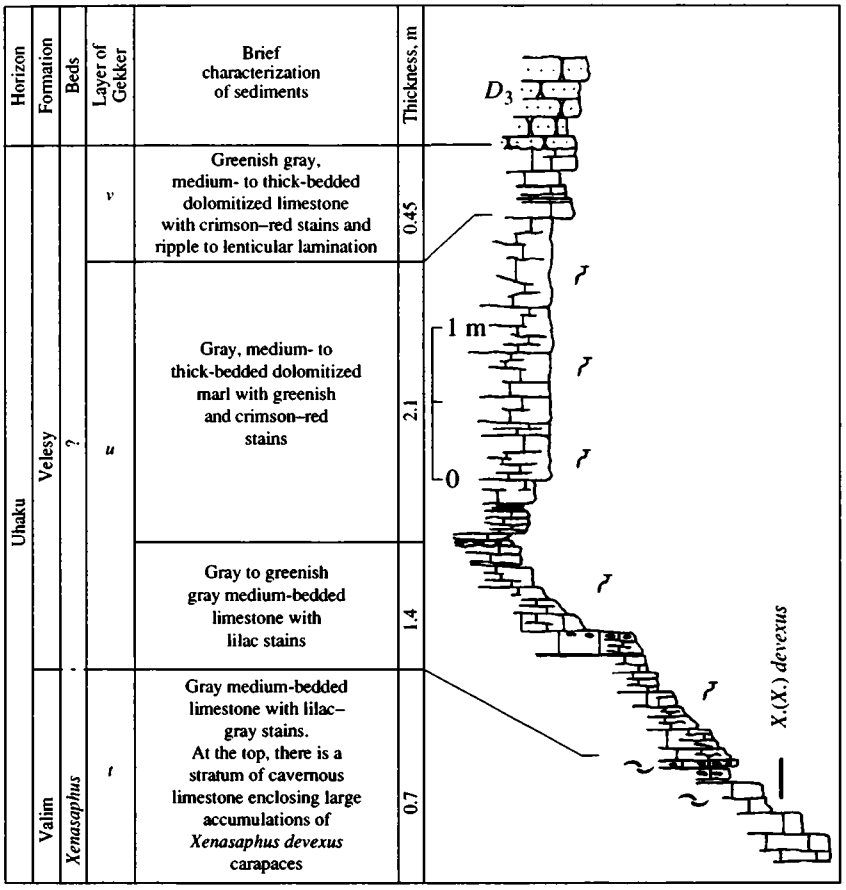

Fig. 6. Outcrop 7, right bank of the Volkhov River downstream of the railway bridge near the Bor Village, the interval between the bridge and nameless creek mouth (symbols as in Fig. 2).

section 6.0 m above the base of the Valim Formation. The range of the beds corresponds to the upper part of the Valim Formation. Species of the subgenus are scarce in the underlying sediments of the Porogi Formation, Lasnamagi Horizon. In the Volkhov River section, one subgenus representative unidentifiable at the species level was found along with *A. bottnicus* 5.10 m above the Porogi Formation base.

The *Chasmops odini–Illaenus intermedius-I. schmidti* Beds conventionally correspond to the Velesy Formation (*Resheniya...*, 1987). The distribution of fauna in the corresponding sections has not been studied thoroughly, and we do not describe it here. In sediments of the Uhaku Horizon, the Volkhov section, ostracodes have not been encountered; their absence is most likely a result of the rocks dolomitization that is unfavorable for preservation of thin shells. The trilobite and ostracode collections are stored at the Paleontological Institute RAS, collection nos. 4330 and no. 4474, respectively.

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DESCRIPTION OF OSTRACODES

The studied ostracodes are mainly represented by known species that are illustrated in Plate III. One new species is described below.

Family Ctenonotellidae E. Schmidt, 1941

Subfamily Steusloffinae Schallreuter, 1966

Genus *Steusloffia* Ulrich et Bassler, 1908

Steusloffia volkhovi Melnikova sp. nov.

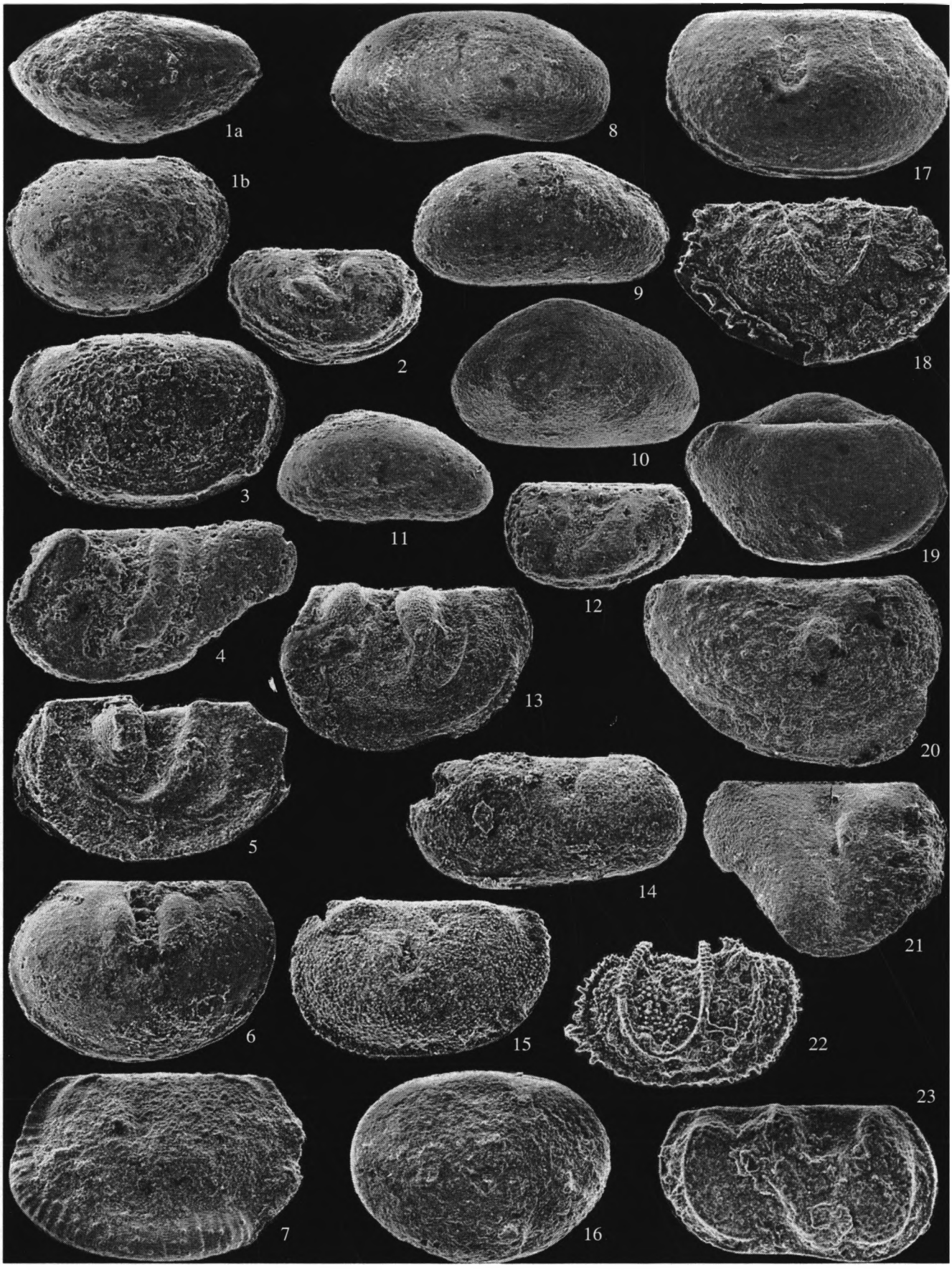
Plate III, fig. 5

The species name is derived from the Volkhov River.

Holotype: PIN no. 4474/ 266, left valve; left bank of the Volkhov River, exposure of the new town bridge (Outcrop 3a); Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon.

Material: 2 shells and 5 valves of various preservation degree.

Description. Shell moderately convex, amplete, unisulcate; S2 is long, small, relatively wide. A large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elevated and rounded middle node L2 with a distinct reticulate surface is developed in front of S2. In front of L2, there is a flattened lobe crossed by crista C1. C1 begins near the anterodorsal angle and crosses the lateral surface towards the lower end of S2. The crista branching point is almost at a mid-height of the valve as viewed from the lower sulcus end. From this point, wide cristas C2 and C3 pass approximately parallel to posteroventral and posterior ends. C2 is more pronounced than C3. The velar structure is represented by a compact ridge extending from the anterodorsal angle to the middle of the posterior end. Shell surface, except for that of finely reticulate L2, is smooth.

Dimensions, mm:

Holotype 4474/266	L = 1.35	H = 0.9
Specimen 4474/267	L = 1.35	H = 0.75

Comparison. The species differs from all known *Steusloffia* forms by wide cristas.

Distribution: Azeri Horizon of the Leningrad region.

Reviewers T.N. Koren' and A.Yu. Rozanov

Plate III. Ostracodes of the Azeri Horizon.

(1) *Leperditella semen* Sarv, 1959; specimen no. 4474/316, shell, L = 2.15 mm, dorsal view (a), side view from the right valve (b); left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Duboviki Formation, 5.35 m from the top of the formation. (2) *Bolbihithis* sp.; specimen no. 4474/336, shell viewed from the right valve, L = 0.7 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (3) *Ahlintella pseudofabuliformis* Sidaravichiene, 1990; specimen no. 4474/326, right valve, side view, L = 0.75 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (4) *Glossomorphites* aff. *acutus* (Hessland, 1949); specimen no. 4474/256, left valve, side view, L = 0.95 mm; left bank of the Volkhov River, Volkhov town, 50–200 m downstream of the VHES dam; Azeri Horizon, Simankovo Formation, 1.40 m above the base of the Azeri Horizon. (5) *Steusloffia volkhovi* Melnikova sp. nov.; holotype no. 4474/266, shell viewed from the left valve, L = 1.35 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (6) *Bollia?* sp.; specimen no. 4474/346, right valve, side view, L = 1.1 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (7) *Eobromidella zvankensis* Meln., 1998; specimen no. 4474/246, shell viewed from the right valve, side view, L = 1.7 mm; right bank of the Koporka River, Kopor'e Village, below the fortress; Azeri Horizon, Duboviki Formation, 1.15 m from the base of the formation. (8) *Longiscula affluensis* Olempska, 1994; specimen no. 4474/301, side view from the right valve, L = 0.95 mm; right bank of the Koporka River, Kopor'e Village, below the fortress; Azeri Horizon, Duboviki Formation, 0.37 m from the base of the formation. (9) *Longiscula parrectis* Neckaja, 1958; specimen no. 4474/306, side view from the right valve, L = 1.2 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (10) *Longiscula arcuaris* Neckaja, 1958; specimen no. 4474/296, side view from the left valve, L = 0.7 mm; right bank of the Koporka River, Kopor'e Village, below the fortress; Azeri Horizon, Duboviki Formation, 0.37 m from the base of the formation. (11) *Longiscula* sp.; specimen no. 4474/311, right valve, side view, L = 1.25 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (12) *Sigmobolbina* sp.; specimen no. 4474/291, left valve, side view, L = 0.9 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (13) *Tallinnella marchica* (Krause, 1889); specimen no. 4474/251, left valve, side view, L = 2.0 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (14) *Primitiella molli* (Bonnema, 1909); specimen no. 4474/341, right valve, side view, L = 0.85 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (15) *Uhakiella cicatriosa* Sarv, 1959; specimen no. 4474/241, left valve, side view, L = 1.35 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (16) *Baltonotella ledai* Sidaravichiene, 1975; specimen no. 4474/321, right valve, side view, L = 1.1 mm; right bank of the Koporka River, Kopor'e Village, below the fortress; Azeri Horizon, Duboviki Formation, 1.15 m from the base of the formation. (17) *Euprimites effusus* Jaanusson, 1957; specimen no. 4474/236, left valve, side view, L = 1.35 mm; right bank of the Koporka River, Kopor'e Village, below the fortress; Azeri Horizon, Duboviki Formation, 1.15 m from the base of the formation. (18) *Piretella oepiki* Thorslund, 1940; specimen no. 4474/276, right valve, side view, L = 1.6 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (19) *Easchmidtella ivantsovi* Melnikova, 1998; specimen no. 4474/261, shell, L = 1.35 mm; right bank of the Koporka River, Kopor'e Village, below the fortress; Azeri Horizon, Duboviki Formation, 1.15 m from the base of the formation. (20) *Gryphiswaldensis* sp. A; specimen no. 4474/231, right valve, side view, L = 1.0 mm; right bank of the Koporka River, Kopor'e Village, below the fortress; Azeri Horizon, Duboviki Formation, 1.15 m from the base of the formation. (21) *Collibolbina plana* (Hessland, 1949); specimen no. 4474/286, right valve, side view, L = 1.05 mm; left bank of the Volkhov River, Volkhov town, 50–200 m downstream the VHES dam, Azeri Horizon, Simankovo Formation, 1.40 m above the base of the Azeri Horizon. (22) *Rakverella (Pectidolon) perornata* (Öpik, 1937); specimen no. 4474/281, left valve, side view, L = 1.45 mm; left bank of the Volkhov River, Volkhov town, exposure of the new automobile bridge, Outcrop 3a; Azeri Horizon, Simankovo Formation, 0.15 m from the base of the Azeri Horizon. (23) *Tetrada (Perplana) perplana* (Neckaja, 1953); specimen no. 4474/271, right valve, side view, L = 0.7 mm; left bank of the Volkhov River, Volkhov town, 50–200 m downstream the VHES dam, Azeri Horizon, Simankovo Formation, 1.40 m above the base of the Azeri Horizon.

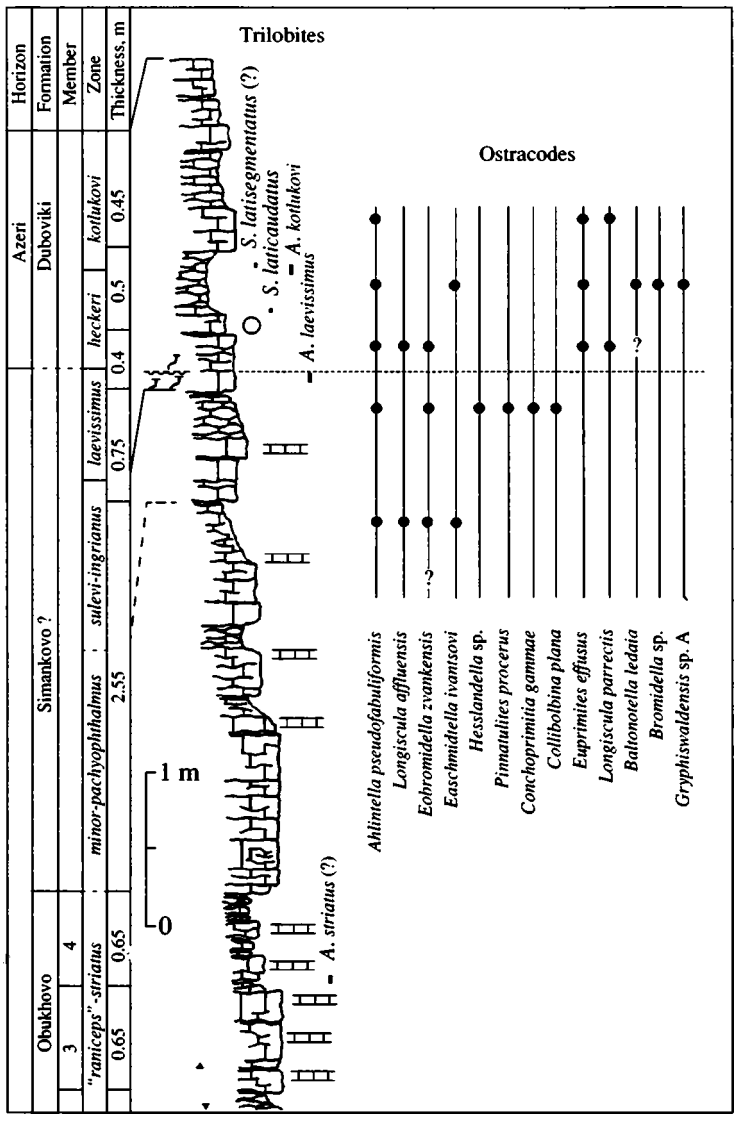

Fig. 7. Right bank of the Koporka River, the site below the fortress Kopor'e Village (symbols as in Fig. 2).

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Upper Paleozoic Stratigraphy in the Eastern Limb of the Tunguska Syneclise

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Abstract—The analyzed materials obtained by editorial preparation of a new map series, scale 1 : 1000000, and generalization of earlier data show that, contrary to prevailing opinions, the Middle to Upper Carboniferous deposits bearing fossil floras, pollen assemblages, brachiopods, and bivalve molluscs of marine, brackish-water, and fresh-water origin are widespread in the eastern Tunguska syneclise. On the contrary, Lower Permian deposits are of extremely limited distribution here and, occurring in the southeast only, they completely wedge out to the northwest. Rare finds of foraminifers, similar to Early Permian arctic forms, do not indicate the Early Permian age of deposits (it would be contradicting data on all other groups of fossils), but suggest presence of facies similar to those, which are peculiar of the Upper Permian in the Arctic basin. Large-leaved cordaitanthaleans most typical of the early Permian occur in association not only with the Late (Alykaevo) but also with Middle Carboniferous (Mazurovo) plant assemblages that testifies to their earlier origin than in Kuzbass.

Key words: volcanogenic formations, Carboniferous, Permian, Tunguska syneclise.

Upper Paleozoic deposits of the Tunguska syneclise eastern limb have been under investigation for a long time, and their paleontological characteristics are known well enough owing to the geological survey and prospecting works carried out here by various organizations of the "Yakutskgeologiya," first of all by the Amakinskaya Expedition. E.M. Vashchenko, N.G. Verbitskaja, L.V. Glukhova, M.V. Durante, I.M. Mashchuk (Tarakanova), S.V. Sukhov, and A.N. Tolstykh identified macrofloras from the study area, palynologists from the "Yakutskgeologiya" and "Irkutskgeologiya" studied palynomorph assemblages, while O.A. Betekhtina and O.V. Lobanova investigated bivalve molluscs. There are however discrepancies and contradictions with respect to the age of deposits. Quite often, both the Carboniferous and Permian plant remains have been found associated in a given stratigraphic unit. Accordingly, some researchers consider it impossible to use plant remains in stratigraphy and operate only with lithostratigraphic subdivisions, which are dated based on individual finds of marine fossils (Chernaya *et al.*, 1999), whereas others explain that many deposits occurring in the region correspond to an "Intermediate Horizon" of Kuzbass (attributed now to the Lower Permian) that yields the mixed assemblages of fossil plants (*Resheniya...*, 1979).

In the 1980s, when a series of new sheets was under preparation for the edition of the Geological map of the USSR, scale 1 : 1000000, the corporation "Aerogeologiya" carried out the inspection and coordination works in upper courses of the Vilyui and Moiero rivers (Fig. 1), within the area of Sheet Q-48, 49 (Aikhal). As for materials concerning subjects of this article, I collected them at that time in collaboration with E.F. Orlova. We obtained new paleontologic materials of a wide significance. Identifying collected plant remains, S.V. Meyen also revised and generalized paleobotanical data known by that time, I.M. Malovetskaja studied fossil bivalves, and V.V. Gershanovich investigated palynological samples. Geological map based on these materials is issued in 1991 (*Geologicheskaya karta...*, 1991), but the explanatory note to it remains unpublished till now. The materials, which have appeared later (Chernaya *et al.*, 1999), are in a good agreement with the former inferences, but some researchers who are not familiar with the analysis done by Meyen suggest erroneously "new" interpretations and call to revision of working stratigraphic schemes. It seems reasonable therefore to publish the necessary generalization of paleontologic materials from the eastern Tunguska syneclise that is the main objective of this work.

CARBONIFEROUS SYSTEM

Carboniferous deposits are represented by gray, greenish gray, and black polymictic sandstones (fine-grained, less often coarse-grained), siltstones, and less abundant shales. These rocks are intercalated with gravelstone and conglomerate interlayers and lenses, and also with coal seams, concretion and thin siderite interlayers. Almost exclusively, these are terrestrial sediments, though their basal portion exposed in the southeastern Alymdzha River basin encloses separate

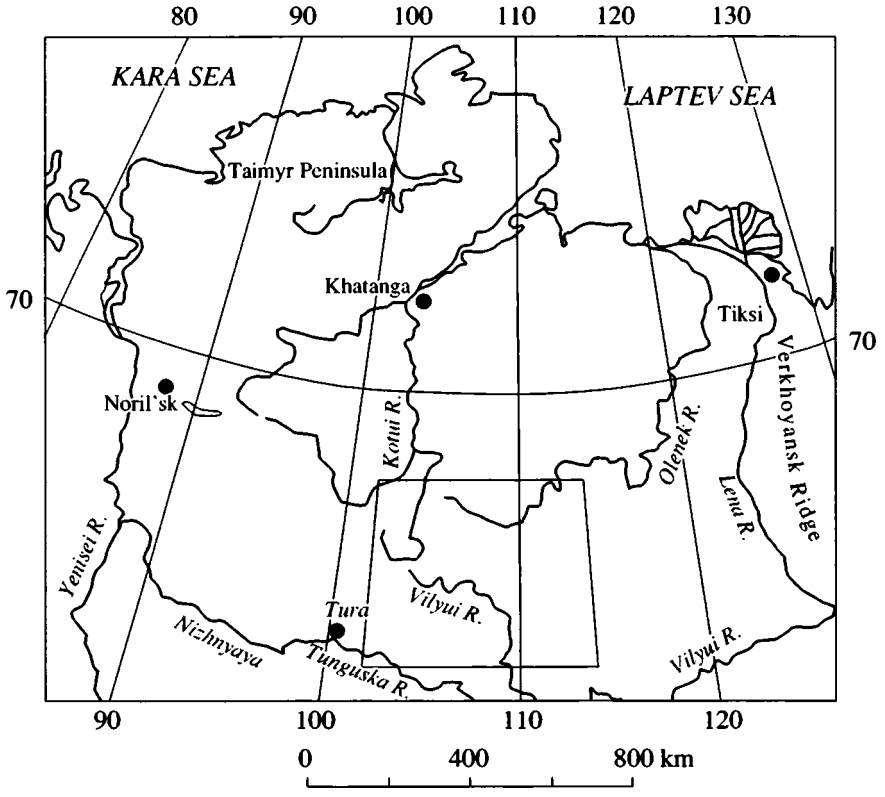

Fig. 1. Schematic geographic map of the study region (straight-line contour outlines map sheet Q-48, 49).

interlayers bearing remains of marine bivalves, foraminifers, and brachiopods.

The deposits overlie with stratigraphic unconformity the various Lower and Middle Paleozoic horizons, from the Lower Ordovician beds in the east and southeast of the region and to the Middle Devonian strata in the northwest. Below them, there is observable the weathering crust that is rather thick (2 m, occasionally up to 5 nm) in the east of the region, e.g., in the Morkoka, Markha, upper Lakharchana, and Alymdzha river basins (Fig. 2). A basal portion of the crust is composed of gravel derived from underlying rocks and cemented by clay. Higher in the section, the crust is represented by brown gravely, sandy, and silty clays or by brown to yellow kaolinite-hydromica and montmorillonite clays, and sometimes (Lakharchana River) by limonitic carbonates.

The Carboniferous assemblages of large-size plant remains occur in many exposures and borehole sections known in the region (Fig. 2). In the boreholes lacking these remains, Permian floral assemblages are established much higher above the base of coal-bearing deposits, and it is impossible to exclude that lower barren intervals of these sections can belong to the Carboniferous. Thus, contrary to prevailing opinions, gray terrigenous deposits of the Middle-Upper Carboniferous seem to be widespread in upper courses of the Moiero, Vilyui and Markha rivers.

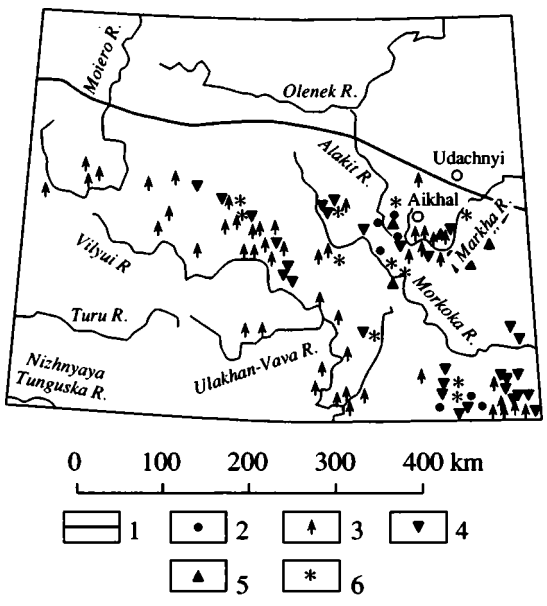

Fig. 2. Localities of Carboniferous paleontological remains: (1) distribution limit of deposits, (2) fossil fauna, (3) fossil flora, (4) palynological assemblages, (5) Carboniferous palynomorphs similar to Permian ones, (6) exposure of pre-Late Paleozoic weathering crust.

Upper Paleozoic stratigraphic scheme for the eastern limb of the Tunguska syneclise

System	Series	Horizon	Structural zones of the Tunguska syneclise		
			Central		Southern
Permian?	Upper	Putorana	Kochechumo Formation Basalts 40–200 m		–
			Nidym Formation Basalts 60–300 m		Limptekon Formation Tuffaceous–sedimentary rocks and tuffs up to 480 m
		Khungtukun	Bugarikta Formation	upper subformation of tuffaceous-sedimentary rocks 60–170 m	
				middle subformation of tuffs 70–100 m	Chichikan Formation Tuffs 70–40 m
		Log Lebedeva		lower subformation of tuffaceous-sedimentary rocks 50–105 m	Yuzhnyi Chun Formation Tuffaceous-sedimentary rocks 120–150 m
		Tutonchana			
Permian	Upper	Gagarii Ostrov	Borulloi Formation Sandstones, siltstones, and shales with coal seams 100–200 m		
		Ust-Degali			
		Pelyatka			
	Lower	Burgukla			Botuoba Formation Sandstones and siltstones 0–50 m
Carboniferous	Middle–Upper	Kata	Aikhal Formation Sandstones, siltstones, and shales with coal seams 10–350 m		
			Konek Formation Conglomerates, gravelstones, and sandstones 0–20 m		Lapchan Formation Conglomerates, gravelstones, and sandstones 0–40 m

The basal Carboniferous strata are almost everywhere composed of conglomerates and gravelstones, or of sandstones with conglomerate and gravely interlayers, being sometimes intercalated with siltstones and shales (Konek and Lapchan formations). These rudaceous strata can be occasionally as thick as 40 m, but usually their thickness does not exceed 5 m and decreases sometimes down to decimeters. The overlying part of the Carboniferous sequence has been distinguished in the study region under the name Aikhal (or Kata) Formation (table). On the Morkoka River left bank, this formation is up to 70 m thick, being much thinner (20–30 m) further to the northeast. In contrast, it is considerably thicker at the right bank of that river and in upper courses of the Vilyui River (120–150 m), being as thick as 230 m in headwaters of the Moiero River and as 350 m in the Vilyui River source area. In the southeast (headwaters of the Olguidakha and Ygyatta rivers), the Carboniferous sequence includes the lower Botuoba Subformation, except for an upper part of the latter in some sections, e.g., in that of Borehole 36. In other sections, where this subformation is of a very low thickness or missing, deposits of the upper Botuoba Subformation also belong to the Carboniferous System (sections of boreholes 5 and 118). In southeastern areas, Carboniferous strata are up to 60 m thick in total. In general, their thickness increases here from 10 to 350 m in the southwestern direction. Most representative sections of deposits under consideration are shown in Fig. 3.

Carboniferous beds yield remains of foraminifers, brachiopods, and bivalves, the latter represented by marine, brackish-water, and lacustrine taxa. They also yield palynomorphs and enclose burials of large-sized plant remains (Fig. 2).

Foraminifers are known from sections of boreholes 27 (Alymdzha River basin, depth interval of 49.6–51.9 m) and Ozernaya 17 (depth interval of 70.6–76.3 m). L.B. Ukharskaya, G.P. Sosipatrova, K.V. Miklukho-Maklay, and L.P. Grozdilova who studied these fossils listed the following their taxa: *Saccammina arctica* Gerke, *Hyperammina borealis* Gerke, *H. rugosa* Gerke, *Hyperamminoides* sp., *Tolypammina* sp., *Trochammina* sp., *Haplophragmoides* sp., *Spiroplectammina* sp., *Cornuspira* sp., *Agathammina* sp., *Glomospira* sp.,

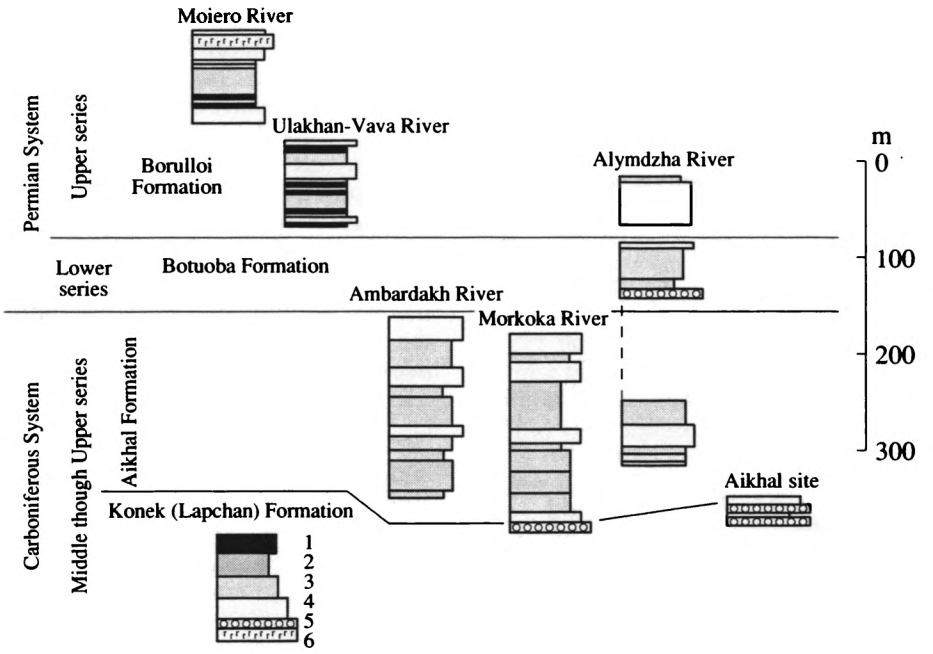

Fig. 3. Correlation scheme of principal Carboniferous-Permian sections (up to Gagarii Ostrov Horizon inclusive): (1) shale; (2) coal; (3) siltstone; (4) sandstone; (5) conglomerate; (6) tuffs and tuffites of basic composition.

Reophax sp., *Endothyra* sp., *Ataxophragmoides tetraxis* Gerke, and Verneuilinidae gen. indet. Coexisting brachiopods from Borehole 27, depth level of 61.5 m, are represented by Carboniferous species *Plicochonetes* aff. *poljenovi* Tolm., *Tomiopsis* sp., and *Neospirifer* cf. *virgatus* (identified by A.D. Grigor'eva and V.I. Ustritskii), and higher in the section, at the depth level of 55 m, A.D. Grigor'eva established presence of the Middle Carboniferous form *Tomiopsis* cf. *tricostata* Kot.

Studying cores from Borehole 27, O.A. Betekhtina and O.V. Lobanova identified Permian marine bivalves within the depth range of 48.8–60.5 m. These are *Streblopteria aclabanensis* Newell, *Myonia* (*Pachymyonia*) aff. *suborbitata* Dick., *Wilkingia komiensis* (Masl.), *Edmondia* sp., *Praeundulomya* (?) sp., *Schizodus* sp., *Pleurophorus* (?) sp., and *Sanguinolithes* sp. In boreholes 23 and 26 situated nearby, they found bivalve species *Abakaniella* cf. *magna* (Tchern.), *Mrassiellina* sp., *Leda* (*Polidevcia*) sp., *Quadratrankulina australiensis* Dick., *Edmondia* sp., and *Praeundulemya* (?) sp. of the same Permian age. A Middle Carboniferous species *Aviculopecten popekoi* (Mur.) was identified by K.A. Astaf'eva in the Aikhal area (Borehole 2005, depth interval of 147–148 m).

The Middle Carboniferous (Mazurovo) assemblages of fresh- to brackish-water bivalves are known from several sites. The most representative one from the Morkoka River basin includes *Kinerkaella elongata* Khalf., *Najadites* sp., *Anthraconaia* (?) cf. *patentis* (Lapsch.), *Mrassiella* ex gr. *magniforma* (Rag.), *Orthonajadites* (?) sp., *Anthraconauta* ex gr. *ragozini* Lapsch., *Abakaniella* cf. *elongata* (Tschern.), and *Procopievskia* (?) sp. (identified by Malovetskaya). Another assemblage of *Abakaniella* cf. *tschernogorskia* (Spassk.), *A.* cf. *jangotoica* (Spassk.), *A.* cf. *lata* (Spassk.), and *Jangotoica plana* Bet. is known from the Aikhal area (identified by Betekhtina from boreholes 0701 and 0901). The Alykaevo assemblages of fresh- to brackish-water bivalves (undivided Carboniferous) are more widespread. Their characteristic forms are as follows: *Kinerkaella* cf. *pseudoedmondia* Bet., *K. balakhonskiensis* (Rag.), *K. pseudobalakhonskiensis* Bet., *K. elongata* Khalf., *K. imitabilis* Khalf., *Mrassiella magniforma* (Rag.), *M.* cf. *ampla* Khalf., *M. ovata* Khalf., *Orthonajadites* cf. *triangularis* Khalf., *Anthraconaia* (?) *vulgaris* Bet., *A.* cf. *fomitchevi* (Fed.), *Anthraconauta* cf. *gracilis* Ivan., *Pseudomodiolus ellipticus* Bet., *Augea* (?) sp. (species from the Morkoka River basin identified by Malovetskaya), *Mrassiellina* cf. *plana* Bet., *M.* cf. *umbonata* (Bet.), *Mrassiella* cf. *magniforma* (Rag.), *M. magniforma* (Rag.) f. *recta* Bet. (ditto, 68 km downstream of the Taseya River mouth; identifications by Betekhtina), *Mrassiella magniforma* (Rag.), *M. rhomboidea* Khalf., *Kinerkaella* aff. *pseudobalakhonskiensis* Bet., *K.* cf. *elongata* Khalf., *Najadites* (?) aff. *kumpani* (Fed.), *Orthonajadites subcentralis* Khalf. (Nizhnii Viluikan site 35 km upstream of the Levyi Viluikan mouth, identifications by Malovetskaya), *Kinerkaella* (?) cf. *pseudoedmondia* Bet., *K.* cf. *ragozini* Bet., *Mrassiellina* cf. *ampla* Khalf., *Abakaniella* (?) cf. *tatarica* Tschern., *A.* (?) *elongata* (Tschern.) (the same site, identifications by Betekhtina), *Angarodon kumsassensis* Rag., and *Mrassiella* cf.

plana Bet. (Borehole 942 in the Aikhal area, identifications by Betekhtina).

Plant remains are even more widespread. Poorly preserved cortex remains of presumably Early Carboniferous *Lophiodendron* sp. are known from the Vilyui River site located 40 km downstream of the Ulakhan-Vava River mouth.[1] The Middle Carboniferous (Mazurovo) floras are reliably established in several sites. For instance, lower siltstone beds exposed in the Sytykan Quarry near the Udachnyi Settlement enclose burials of *Angarodendron obrutschevii* Zal. (abundant), *Angaridium finale* Neub., *A. mongolicum* Zal., *Angaropteridium cardiopteroides* (Schmal.) (abundant), *Rufloria archaica* Gluch., *R. subangusta* (Zal.) (abundant), *R.* ex gr. *derzavinii* (Neub.) vel *subangusta* (Zal.) (abundant), *Cordaites* aff. *singularis* (Neub.), cf. *Gaussia cristata* Neub., *Krylovia sibirica* Chachl., and *Samaropsis moracia* Zal. Similar plant assemblages are known in seven other sites of the region. In addition to taxa listed above, they include *Angarophloios planus* (Neub.), *Koretrophyllites mungaticus* Radcz., *Paracalamites costatus* Gorel., *Angaridium potaninii* (Schmal.), *Neuropteris tomiensis* (Zal.), *N. izylensis* (Tchirk.), *Cordaites indeterminatus* Gluch., *C. neuburgii* Gluch., *Rufloria permulta* Gluch., *R. theodorii* (Tchirk. et Zal.), *R.* ex gr. *meyenii* Gluch., *Ginkgophyllum vsevolodii* Zal., *Cardiocarpus krivljakensis* Such., *Samaropsis pauxilla* Zal., *S.* (?) *angarica* Rassk., *S. auriculata* Neub., *Angarocarpus ungensis* (Zal.), and *Holcospermum* (?) *tcheltchetensis* Such. (identifications by Vashchenko, Glukhova, Meyen, and Sukhov). It is important that the Middle Carboniferous (Mazurovo) species coexist in this assemblage with rather abundant taxa (mostly of Cordaitales), which appear in Kuzbass beginning since the "Intermediate Horizon" only and are abundant in the Lower Permian (upper Balakhon) deposits. Exactly these taxa define the Early Permian age of the horizon, and their admixture was thought to be characteristic of the Late Carboniferous floras in the study region. Analyzing regional data by edition of the sheet Aikhal of geological map (*Geologicheskaya karta...*, 1991), Meyen wrote in 1984: "...as is established..., cordaitalean leaves from a rather low level of regional sequence, i.e., from that correlative with the Mazurovo Horizon of Kuzbass, have microstructure that is typical of Cordaitales occurring there much higher, in the upper Alykaevo strata and Intermediate Horizon. Sometimes these "younger" Cordaitales are even dominants. Consequently, it is necessary to reexamine these plants. They may have diagnostic features appropriate for discrimination between these and true young forms. This is one of the keen problems in Upper Paleozoic phytostratigraphy of Siberia. I think that proper study may also result in recalibration of the age level occupying by Siberian floras in the International Stratigraphic Scale. This level is actually defined based on relations between marine and terrestrial strata in the Verkhoyansk and Taimyr regions, first of all in the former one. The traditional viewpoints are inconsistent with situation elucidated now in the east Siberian platform. In the Verkhoyansk region, the intermediate flora (= Klintaiga Horizon) is associated with marine fauna of the Upper Carboniferous. That is why M.V. Durante concluded that the Carboniferous–Permian Boundary is inside this horizon. However, it seems admissible now to think that the Verkhoyansk flora resembling plant remains of the intermediate type is actually older, like in the East Siberian platform, than these remains."

Most widespread in the region are the Late Carboniferous (Alykaevo) assemblages of plant fossils. In addition to some aforementioned species of a wide stratigraphic range, these assemblages include *Paragondwanidium sibiricum* Pet., *P. odontopteroides* (Zal.), *Pursongia asiatica* Zal., *Zamiopteris* ex gr. *glossopteroides* Schmal., *Samaropsis patula* Zal., *Samarospadix penicellata* Zal., *Cordaicarpus kovbassinae* Such. and some other forms (identifications by Meyen and Durante). In three various sites of the region, there were encountered floral assemblages, which may belong to either the uppermost Carboniferous, or the lowermost Permian. In these cases, the species listed above are associated with *Rufloria* ex gr. *barzassica* Gluch., *R.* ex gr. *birjulinkiensis* Gluch., *R.* ex gr. *poryvaica* Gluch., *R.* ex gr. *khalfinii* Gluch., and *Bardocarpus discretus* Neub., all unknown below the uppermost Alykaevo strata.

Palynological assemblages of Carboniferous deposits are quite variable in composition. Usually they are dominated by spores and less frequently by pollen. Taxa most frequent among spores are *Cyclobaculisporites trichacantus* (Lub.) and other tuberculate forms (10–55%), *Remysporites psilopterus* (Lub.), and *R. mirabilis* (Lub.) (4–50%), though their abundance can be very low sometimes. Less abundant species (up to 15–20%) are *Apiculatisporites spinosus* (Naum.), *A. hispidus* (Andr.), *Nigrisporites nigritellus* (Lub.), and *N. nigrotuberculatus* (Lub.), though in places they may show a peak abundance. Quite common spore forms are *Lycospora breviapiculata* (Lub.), *L. subtriquetra* (Lub.), *L. echinata* (Andr.), and *L. verriculifer* (Lub.) representing 10 to 15%, occasionally 32% of the whole spectrum. Almost constantly occurring *Verrucosisporites ermakovianus* (Koval.) is of a low abundance everywhere. *Capillatisporites lunatus* (Rusk.), *Cyclogranisporites larvatus* (Lub.), *C. polipyrenus* (Lub.), *Calamospora microrugosa* (Ibr.), *Punctatisporites, Turrisporites rigidispinosus* (Lub.), *T. pyramidalis* (Lub.), *Leiotriletes extensus* (Lub.), and *Spinosisporites* spp. represent frequent, though not abundant spore species. Some spectra reveal presence of *Raistrickia* forms, which are abundant sometimes.

Cordaitanthaleans *Cordaitina rotata* (Lub.), *C. uralensis* (Lub.), *Guthoerlisporites verus* (Sadk.), *G.*

[1] If not specified otherwise, plant taxa are given here in systematics by S.V. Meyen.

varians (Sadk.), *Libumella rugulifera* (Lub.), *Luberisaccites stipticus* (Lub.), and *L. subrotatus* (Lub.), which are dominant in the pollen group, associate rather regularly with *Cordaitina minor* (Lub.), *C. angustelimbata* (Lub.), *Samoilovitchisaccites turboreticulatus* (Samoil.), and *Crucisaccites ornatus* (Samoil.). Grains of *Florinites* spp. occur almost everywhere. They represent few to 14% of the pollen spectra and locally may be absent. In the Aikhal area, their abundance in lower beds of the Carboniferous sequence is as high as 32%.

The palynological assemblages described above and having an insignificant content of pollen are undoubtedly of the Carboniferous age. Those of them, which are rich in *Lycospora* and *Florinites* species and include some other older taxa, can be attributed to the Middle Carboniferous (Mazurovo Horizon), while the others, which reveal either an insignificant content or complete absence of these components, are of the Late Carboniferous age (Alykaevo Horizon). The former have been encountered, for instance, in the Alymdzha River basin (Borehole 27, depth interval of 50.7–54.2 m). Predominance of cordaitanthalean pollen used to be considered as a diagnostic feature of the Early Permian palynological assemblages, and admixture of Carboniferous miospores was commonly regarded in such a case as a result of their redeposition. Nevertheless, predominance of cordaitanthalean pollen has been established in several outcrops of the upper Aikhal Subformation in the eponymous area. Consequently, palynomorph assemblages of this kind may be characteristic, like burials of macrofloras with abundant cordaitanthalean leaves of the Permian habit, of the Carboniferous deposits.

Finally, there are known Carboniferous palynological assemblages of the following composition: smooth (12 to 40%), acanthous (2 to 26%), and echinate to tuberculate (4 to 16%) spores in association with cordaitanthalean (13 to 41%) and *Entylissa* (8 to 20%) pollen. Characteristic of these assemblages are *Nigrisporites nigritellus* (Lub.), *N. marginatus* (Portn.), *Leiotriletes triangulatus* (Andr.), *L. extensus* (Lub.), *Turrisporites sibiricus* (Medv.), *Raistrickia cornuta* (Andr.), *R. heteromorpha* (Andr.), *R. obtusosaetosa* (Lub.), *R. grandispinosa* (Andr.), *R. multicolorius* (Andr.), *Cordaitina minor* (Lub.), *Libumella rugulifera* (Lub.), *Entylissa caperata* (Lub.), *E. glabra* (Andr.), and some other taxa. Carboniferous miospores always present in the assemblages represent 5 to 50% of their composition. Microfloras under consideration used to be dated back to the Late Permian, and their Carboniferous palynomorphs were regarded as redeposited. However in several localities, e.g., in lower courses of the Markha (Borehole 5), Alakit (Borehole 0907), and Alymdzha (Borehole 37) rivers, or near junction of the Lakharchana and Khakhsyk rivers, they occur in association with leaf remains of Carboniferous plants.

From the formal standpoint, deposits bearing that peculiar association of micro- and macroflora can be regarded as spanning a wide stratigraphic interval, from the Lower Carboniferous to Upper Permian, inclusive. The proper analysis of available materials leads, however, to the next conclusions. The Early Carboniferous age admissible for their lower beds cannot be taken for granted. A greater interval of the sequence can be attributed quite reliably to the Middle-Upper Carboniferous. Based on composition of floral remains, it impossible either to exclude, or to prove the Early Permian age of upper beds in the sequence. The lower sequence part cannot be referred, based on marine bivalves and foraminifers occurring here, to the Lower Permian, as it would be contradicting paleobotanical data. Paleobotanical characterization of this interval is well substantiated, whereas foraminifers and marine bivalves are known from a single borehole and any data on their evolution in the basin are unknown. It is possible to assume as well that this borehole penetrated a subvertical zone of tectonic displacement (faulting of this kind can be frequently observed in the Tunguska syneclise), and then the lower portion of penetrated sequence could be younger than the upper one. Finally, it is impossible to argue, based on palynological data, for the Late Permian age of uppermost strata, because pollen of that age occur in association with the Carboniferous macroflora.

PERMIAN SYSTEM

Gray terrigenous deposits of the Permian System, which are similar in composition to Carboniferous strata, are widespread in the Moiero, Vilyui, Alymdzha, Olguidakh, and Ygyatta riverheads. Their sequence is divisible into the lower and upper series.

Lower Series

The Lower Permian strata constituting a part of the Botuoba Formation sequence are well studied in the southeast of the region, in the Olguidakh and Ygyatta river basins (Fig. 4). They are composed of gray to black sandstones, siltstones, and shales intercalated with gravelstone and conglomerate interlayers. Basal Permian beds usually overlie Carboniferous deposits, though in the east they may locally rest directly on the Ordovician and Silurian sediments (*Geologicheskaya karta...*, 1991). In the last situation, the Permian age of beds above the latter has not been substantiated by paleontological data, and I do not exclude that the overlying sequence actually begins with thin Carboniferous layers. The Lower Permian sequence is up to 50 m thick and has been most completely cored in Borehole 116 (Figs. 3 and 4).

Remains of Permian plants have been encountered in several depth intervals penetrated by boreholes 4, 36, and 116 in the Olguidakh River basin. These are *Paracalamites* cf. *planicostatus* Verb., *P. prokopievskiensis*

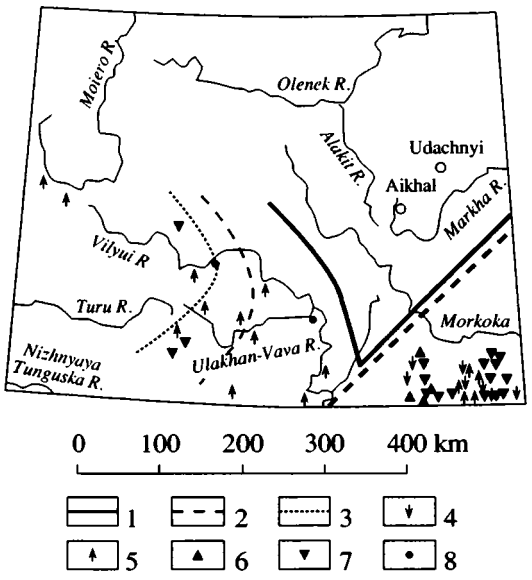

Fig. 4. Localities of Permian paleontological remains: (1) boundary between structural zones; distribution areas of Degali (2) and Gagarii Ostrov (3) formations; (4) Early and (5) Late Permian floras; (6) Early and (7) Late Permian palynological assemblages; (8) faunal remains (other symbols as in Fig. 2).

Gorel., *P. vicinalis* Radcz., *P. costatus* Gorel., *Annularia* cf. *tenuifolia* Neub., *Rufloria* cf. *derzavinii* (Neub.), *R.* cf. *tajmyrica* (Schved.), *R. theodorii* (Tchirk. et Zal.), *R. kryschtofovitschii* (Radcz.), *R. tchirkovae* (Zal.), *Cordaites magnus* (Schved.), *C. latifolius* (Neub.), *Crassinervia tunguskana* Schved., and *Evenkiella shortonotensis* Gorel. (identified by Verbitskaya and Mashchuk-Tarakanova).

Pollen (up to 66%) is a dominant component of corresponding palynological assemblages. It represents almost exclusively *Cordaitina rotata* (Lub.) whose grains infrequently associate with pollen of *C. minor* (Lub.), *C. angustelimbata* (Lub.), *Guthoerlisporites varians* (Sadk.), *Samoilovitchisaccites turboreticulatus* (Samoil.), and *Crucisaccites ornatus* (Samoil.). Also abundant are smooth spores. Rather frequent (up to 20%) pollen of the genus *Apiculatisporites* is represented by species *A. rectispinus* (Lub.), *A. parvispinus* (Lub.), *A. alligodontus* (Andr.), *A. exacutus* (Andr.), *A. multangulus* (Andr.), *A. spine* (Naum.), and *A. hispidus* (Andr.). Other characteristic taxa are abundant *Raistrickia cornuta* and less frequent *Granulatisporites piriformis* Loose, *Nigrisporites lemniscatus* (Lub.), *N. nigritellus* (Lub.), *N. nigrotuberculatus* (Lub.), and *Cyclogranisporites polipyrenus* (Lub.). Some palynological spectra, e.g., that from the depth level of 48 m in Borehole 4, are depleted in pollen, being enriched in Carboniferous spores, redeposition of which used to be assumed.

In the Unified Stratigraphic Scheme (*Resheniya tret'ego...*, 1982), the entire Botuoba Formation is attributed to the Lower Permian. The analyzed paleobotanical data on the Olguidakh and Ygyatta river basins imply, however, that the Lower Permian corresponds only to the upper lower and upper Botuoba subformations (except for section of Borehole 5, in which Vashchenko identified plant remains of indisputable Carboniferous age). Plant taxa characteristic of this stratigraphic interval are *Rufloria* cf. *pseudoaequalis* (Radcz.) *R.* cf. *derzavinii* (Neub.), *R. theodorii* (Tchirk. et Zal.), *Cordaicarpus rotundatus* (Radcz.), *Samaropsis skokii* (Neub.), *S.* cf. *patula* Zal., and *Bardocarpus tunguskiensis* Verb. (identified by Mashchuk). Single specimens of *Sphenophyllum prokopievskiensis* Gorel., *Rufloria* ex gr. *derzavinii* (Neub.), *Nephropsis rhomboidea* Neub., and *Samaropsis danilovii* Such. are known from boreholes 0452, 568, 570, 0754, and 0854 drilled in the Aikhal area. Palynological assemblages macerated from core samples of three boreholes (358, depth interval of 7–21 m; 431, depth interval of 41–66 m; 1224, depth interval of 18–28 m) contain abundant (up to 50%) and diverse pollen of Cordaitales. Species identified among the latter are *Crucisaccites ornatus* (Samoil.), *Cordaitina rotata* (Lub.), *C. uralensis* (Lub.), *C. angustelimbata* (Lub.), *C. undata* (Lub.), *C. abutiloida* (Andr.), *C. minor* (Lub.), *Luberisaccites stipticus* (Lub.), *Libumella rugulifera* (Lub.), *Guthoerlisporites varians* (Sadk.), *G. verus* (Sadk.), and *Samoilovitchisaccites turboreticulatus* (Samoil.). They associate with abundant smooth, tuberculate, and echinate spores representing *Nigrisporites nigritellus* (Lub.), *N. nigrotuberculatus* (Lub.), *Granulatisporites facerus* (Andr.), *G. piriformis* Loose, *Apiculatisporites spinosus* (Naum.), *A. hispidus* (Andr.), *Spinosisporites rectispinus* (Lub.), *Raistrickia obtusosaetosa* (Lub.), and *R. cornuta* (Andr.). Spore species *Cyclobaculisporites trichacantus* (Lub.), *Remysporites psilopterus* (Lub.), and those representing genera *Protohaploxypinus, Protodiploxypinus*, and *Lueckisporites* are much less abundant. The last genus known in Russia from the Tatarian Stage and, presumably, from the upper Kazanian Substage is characteristic of Lower Permian deposits in Western Europe. In all other aspects, palynomorphs listed above are typical of the Lower Permian. Their association with plant remains in the Aikhal area may indicate that Lower Permian deposits occur also outside the Olguidakh and Ygyatta river basins. Assuming this, one should take into account two remarks. First, it is still not undisputable that these associations are of the Permian age, and similar combinations of leaf remains and miospores occasionally occur in Carboniferous strata. Second, Permian deposits, if any, may occur as small outliers in the aforementioned area, being of an insignificant thickness. In addition, identifications of marine bivalves *Kolymia* cf. *irregularis* Lich. and *K.* ex gr. *pterinaeformis* Lich., which have been reported from the junction area of the Ulakhan-Vava and Kyuryungnekyan rivers, should be accepted critically, as they seem to be grounded by poorly preserved specimens.

It is clear from the aforesaid why the Lower Permian deposits are not shown in the geological map of the region scaled down to 1 : 1000000, although their relic beds may be present here. Rasskazova (1958) was first to note that Lower Permian strata are missing from sedimentary sequence of the eastern Tunguska basin.

Upper Series

The Upper Permian deposits are widespread in upper courses of the Moiero, Vilyui, Olguidakh, and Ygyatta rivers (Fig. 5). Further northeastward, in the Levyi Nizhnii Vilyuikan, Mogdy, Dzheltuli, Khakhsyk, Lakharchana, Morkoka, and Markha river basins, their presence, except for local outliers, has not been established without doubts. The Upper Permian sequence is composed of gray to greenish gray and black sandstones (usually fine-grained) intercalated with siltstone or shale interlayers and enclosing thin coal seams. Interlayers (up to 0.9 m thick) of tuffite, tuffaceous sandstone, and ashy tuff are characteristic of the upper part (12 m) of the sequence exposed in the Moiero riverhead, and interbeds of bituminous siltstone are present in exposures of the Eika River. The Upper Permian strata discordantly overlie either the Lower Permian, or Carboniferous deposits, and occur above Ordovician or Silurian sediments in eastern areas of the region, where Permian age of their basal portion has not been proved by paleontological data. The Upper Permian sequence is 100 to 200 m thick. Most representative sections of this stratigraphic interval are illustrated in Fig. 3.

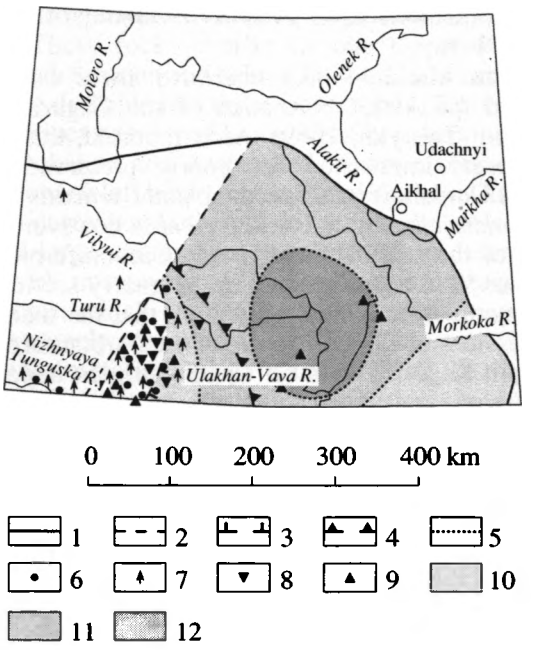

Fig. 5. Localities of Late (?) Permian paleontological remains: (1) distribution limit of deposits; (2) boundary between central and Southern structural zones; (3) distribution limit of Pirda sequence; (4) distribution limit of Limptekon Formation; (5) boundary between facies of Yuzhnyi Chun Formation; (6) fossil fauna; (7) fossil flora; (8) Mesophytic and (9) Paleophytic palynological assemblages; (10) distal ashy tuffs and tuffites, (11) intercalated distal and proximal (coarse tuffs) facies, and (12) proximal coarse tuffs of the Yuzhnyi Chun Formation.

In the Olguidakh and Ygyatta river basins (boreholes 33, 39, 117, 118), Upper Permian strata bear remains of *Equisetina* cf. *tenuistriata* Radcz., *Annularia batchatensis* (Chachl.), *A.* aff. *rarifolia* Radcz., *Paracalamites* aff. *grandis* Gorel., *Petscheria* aff. *oblonga* Gorel., *Cordaites* cf. *candalepensis* (Zal.), *C.* cf. *concinnus* (Radcz.), *C.* aff. *oblongatus* (Radcz.), *C.* cf. *kusnetskianus* (Gorel.), *Rufloria* cf. *olzerassica* (Gorel.), *Crassinervia minima* Such., *C.* cf. *pentagonata* (Gorel.), *Nephropsis* aff. *schmalhausenii* Radcz., *Cordaicarpus tagarischskiensis* Such., *C. iljinskiensis* Radcz., *C. petrikensis* Such., *Samaropsis* aff. *pseudotriquetra* Neub., and *Skokia oblongata Such.*, (identified by Mashchuk and Verbitskaya). This assemblage is most likely characteristic of the basal Upper Permian. *Rufloria aperta* Gluch., *R.* cf. *brevifolia* (Gorel.), and *Cordaites* ex gr. *gracilentus* (Gorel.) known from the Ambardakh and Vavukan river basins characterize there the Pelyatka Horizon of the basal Upper Permian. In the Ulakhan-Vava river basin, fossil flora of the Degali (Ust-Degali) Horizon is represented by *Tychtopteris* sp., *Cordaites gracilentus* (Gorel.), *C. clercii* (Zal.), *C. adleri* (Radcz.), *Lepeophyllum* cf. *belovaensis* Gorel., *Crassinervia* sp., *Niazonaria* sp., and *Tungussocarpus tychtensis* (Zal.). In the Vilyui River sites and in the Moiero riverhead, there is known the Gagarii Ostrov assemblage of *Annularia jerunakovensis* Neub., *Phyllotheca turnaensis* Gorel., *Pecopteris* cf. *anthriscifolia* (Goepp.), *P.* cf. *compta* (Radcz.), *P. leninskiensis* (Chachl.), *Cordaites insignis* (Radcz.), *C. clercii* (Zal.), *C, adleri* (Radcz.), *C. angustifolius* (Neub.), *N. khalfinii* (Gorel.), *C. oblongifolius* (Radcz.), *C. minimus* Neub., and *Lepeophyllum* sp.

Smooth (15 to 60%) and echinate (2 to 64%) spores prevailing in palynological assemblages are associated with less abundant tuberculate spores (6 to 30%) and with cordaitanthalean (6 to 60%) and monocolpate (1–7 to 40% in uppermost strata) pollen. Most characteristic palynomorph taxa are *Nigrisporites nigritellus* (Lub.), *N. marginatus* (Portn.), *N. nigrotuberculatus* (Lub.), *Leiotriletes extensus* (Lub.), *L. ternaries* Andr., *L. triangulus* (Andr.), *Turrisporites sibiricus* (Medv.) (up to 20%), *Spinosisporites parvispinus* (Lub.), *S. acinaciformis* (Andr.), *Granisporites acutus* (Andr.), *Raistrickia cornuta* (Andr.), *R. heteromorpha* (Andr.), *R. obtusosaetosa* (Lub.), *R. grandispinosa* (Andr.), *R. abrupta* Andr., *R. papillaris* (Andr.), *Cordaitina minor* (Lub.), *C. angustilimbata* (Lub.), *C. rotata* (Lub.), *Entylissa caperata* (Lub.), and *E. glabra* (Andr.). Associated Carboniferous miospores occur as single grains.

Thus, the Upper Permian deposits are represented in the region by different horizons, the younger of which

occur in the axial zone of syneclise, being of a limited distribution.

In the Markha-Alakit interfluve, there was distinguished the Alakit Formation of the Gagarii Ostrov Horizon (Tolstykh, 1969). As is reported, this formation bears impressions of poorly preserved small-leaved Cordaitales and peculiar plant remains defined as *Markhajella*. In the stratotype area however (northward of the Lake Kyllakh), only Carboniferous plant remains have been detected in the quarry sections and core samples. It is likely therefore that the mentioned impressions of Cordaitales are of the Carboniferous age as well. Reported data on Upper Permian deposits in the Aikhal area cannot be regarded as properly proved.

UPPERMOST PERMIAN–BASAL TRIASSIC

In the Vilyui, Moiero, Markha, and Alakit riverheads and along the Nizhnyaya Tunguska riversides, there are widespread volcanogenic rocks of basic composition. Their sequences are of variable structures in different areas belonging to Central and Southern zones of the Tunguska structural–lithologic region (*Resheniya tret'ego...*, 1981; *Geologicheskaya karta...*, 1991). Boundary between these zones approximately corresponds to the line connecting mouths of the Nizhnii Kochechumo and Srednii Vilyuikan rivers (Fig. 5).

Central Structural Zone

The Central structural zone includes a part of the Nizhnyaya Tunguska River valley (in its middle courses), Nizhnii Kochechumo and Korvuchana river basins, Turu middle courses, and Moiero and Vilyui riverheads. The volcanogenic sequence is composed here of tuffaceous to sedimentary rocks of the Bugarikta Formation (Korvuchana Group) in its lower part and of basaltic lavas and volcanogenic-sedimentary rocks of the Nidym and Kochechumo formations of the Putorana Group. Not far away from distribution limits of the Putorana Group, there is established boundary separating different-rank coals of the Gagarii Ostrov Formation (the high-volatile from bituminous ones), and as is suggested, lava flows formerly advanced close to this boundary.

Rocks of the Bugarikta Formation are exposed in the Korvuchana, Turu (middle courses), Moiero, Vilyui (upper streams), and Nizhnyaya Tunguska river basins. These are green to brown, sometimes gray to yellowish and reddish beds of tuffaceous siltstones and sandstones, tuffites, tuffs (xenotopic varieties inclusive), and volcanomictic rocks. The formation is divided into three subformations interconnected via gradual transitions.

In the Moiero River basin, the lower Bugarikta Subformation (= Eksa Formation in the western syneclise limb, Sadovnikov and Orlova, 1995) discordantly overlies the Upper Permian coal measures. In outcrops along the Nizhnyaya Tunguska River, its basal beds are unexposed. The unit is composed of layered volcanogenic–sedimentary rocks, mainly of ashy tuffs and tuffites intercalated with interlayers of coarse tuff in the upper part and with basaltic lava flows in the lower one. The total thickness of subformation is up to 105 m.

Conchostracans from the lower Bugarikta Subformation are represented by species *Rohdendorfium (Bipemphigus) tutontchanum* (Nov.), *Sphaerograpta deshaseauxae* Nov., *S. feysi* Nov., *Tripemphigus sibiricus* Nov., *Sphaerestheria sibirica* Nov., and *Falsisca secreta* Nov. Concurrent ostracode species (Nizhnyaya Tunguska sites) are *Darwinula* cf. *mera* Misch., *D. postparallela* Misch., *D. regia* Misch., *Suchonella lucida* Mand., *Gerdalia* aff. *clara* Misch., *G. variabilis* Misch. Plant remains known from the subformation are those of *Equisetum* (?) cf. *triassicum* (Radcz.), *Todites polymorpha* (Mog.), *Osmundopsis* (?) ex gr. *uralica* (Pryn.), *Cladophlebis* cf. *augusta* Heer (Moiero sites), *Edyndella* (?) sp., and *Quadrocladus* cf. *sibiricum* (Neub.) (Nizhnyaya Tunguska sites). Fossils listed above imply that the unit under consideration can be attributed to the terminal Permian Khungtukun Horizon of the Taimyrian regional stage (Sadovnikov and Orlova, 1994).

The middle Bugarikta Subformation of coarse tuffs with interlayers of ashy tuffs, tuffites, and volcanogenic–sedimentary rocks is an equivalent of the Kholokit Formation distinguished in the southwestern limb of the Tunguska syneclise (Sadovnikov and Orlova, 1995). These rocks overlie either the lower subformation, or the Upper Permian coal measures. Their total thickness is 100 to 250 m (occasionally up to 400 m). Fossils known from this subdivision are fish remains *Colobodus* (?) and *Tungusichtys*.

The upper Bugarikta Subformation, an analogue of the Irgakta Formation in the southwest of the syneclise (Sadovnikov and Orlova, 1995), rests on the erosion surface of the lower and middle subformations. It is composed of volcanogenic–sedimentary and volcanomictic rocks enclosing interlayers of ashy to coarse tuffs. The unit is 60 to 100 m thick (170 m at a maximum).

Fossils of the upper Bugarikta Subformation are known from the Nizhnyaya Tunguska sites. These are *Rohdendorfium (Bipemphigus) gennisi* (Nov.) and *Megasitum lundongaensis* Nov. representing conchostracans and plant remains *Acrostichides tunguskanus* (Pryn.), *Mertensides lingulates* Mog., *Cladophlebis* cf. *crenata* Pryn., *C.* cf. *kaoiana* Sze, *C. kirjamkensis* (Pryn.), *C. lobifera* Pryn., *Parajacutiella* (?) *angusta* Mog., and *Elatocladus* (?) sp. They imply that their host deposits correspond to the terminal Permian Khungtukun Horizon of the Taimyrian regional stage (Sadovnikov and Orlova, 1994).

Rocks of the Nidym Formation are widespread in the Moiero (riverhead areas), Korvuchana, Tala, Turu (beyond riverhead areas), Nizhnii Kochechumo, and Nizhnyaya Tunguska river basins, where they overlie

the upper Bugarikta Subformation. These are thin lava flows (locally pillowed) of dark gray fine-grained basalts with distinct amygdules and cindery crust at their surface, which are intercalated with interlayers and lenses of brown to reddish tuffs and tuffites. In the right bank of the Nyukchorok River, fine-grained basalts may represent anamesite sills (?). The Nidym Formation is usually as thick as 60 to 140 m, being thicker only in the Turu River basin (200 m) and in the Nyukchorok-Moiero interfluve (300 m). In nearby region, the formation yields floral and faunal remains characteristic of the terminal Permian Khungtukun Horizon (Sadovnikov and Orlova, 1994).

The Kochechumo Formation of basaltic lavas with intercalations of grayish green to brown and reddish volcanogenic–sedimentary and volcanomictic rocks is distinguishable in the Korvuchana, Turu (beyond riverhead area), Nizhnii Kochechumo, and Nizhnyaya Tunguska river basins. It is divided into three subformations. The lower one, an analogue of the Agitkan Formation in the axial syneclise zone (Sadovnikov and Orlova, 1995), conformably overlies the Nidym Formation. Volcanogenic-sedimentary and volcanomictic rocks of this subdivision (40 to 200 m thick) enclose scarce lava flows (one or two) of fine-grained basalts displaying fine prismatic jointing (anamesites ?). Fauna and flora characteristic of the Putorana Horizon (Taimyrian regional stage of the terminal Permian) was found in correlative rocks exposed in neighboring regions. The middle Kochechumo Subformation, an equivalent of the Kondakan Formation in the syneclise axial zone (Sadovnikov and Orlova, 1995), conformably overlies the lower one and is composed of thin dark gray lava flows of amygdaloidal and pillowed fine-grained basalts (110 to 180 m thick in total). The upper Kochechumo Subformation is of a limited distribution in some watersheds, e.g., between the Lenko and Verkhnii Turukan rivers, the Turu River basin. This subdivision conformably resting on the middle one is composed of thick lava flows of dark gray massive basalts lacking amygdules (170 m thick in total).

Southern Structural Zone

The Kananda, Eika, Turu (riverhead area), Vilyui (beyond its riverhead), Morkoka, and Markha river basins belong to the Southern structural zone. The volcanogenic Yuzhnyi Chun, Chichikan, and Limptekon formations distinguished here are composed almost exclusively of volcaniclastic and tuffaceous sedimentary rocks. In the east, the last formation has not been recognized, and rocks of two others are of a limited thickness and occupy insignificant areas in the Morkoka, Markha, Lakharchana, Alymdzha, Olguidakh, and Ygyatta river basins.

Rocks of the Yuzhnyi Chun Formation are exposed in the Vilyui (upstream of the Srednii Vilyuikan mouth) and Ulakhan-Vava river valleys. Further eastward, they are recognizable in separate exposures along the Morkoka, Markha, Alymdzha, Olguidakh, and Ygyatta rivers. These rocks overlie various Upper Paleozoic strata and Silurian to Ordovician horizons in the east. They are divisible into three facies types (Fig. 6). In the Vilyui River basin (right-side areas between mouths of the Vavukakan and Ambardakh rivers, except for the Lakharchana left bank), rocks of the Yuzhnyi Chun Formation correspond to proximal facies, i.e., to coarsely bedded large- to medium-grained tuffs and xenotopic tuffs with interlayers of ashy tuffs and volcaniclastic sediments, which may be frequent in some exposures and extremely scarce in the others. In sections of this kind, the formation is 120 to 150 m thick. To the west, northeast, and east of their distribution area, in the Vavukan, Ulakhan-Vava (middle courses), Eika, Betere, Morkoka (between Taseya and Tangkhaya rivers), and Markha (head area) river basins, coarse-grained tiffs constituting the middle interval (about 40 m thick) are underlain and crowned by beds of distal layered tuffites and volcano-sedimentary rocks. In this case, the lower and upper members are up to 35 and 85 m thick, respectively, and total thickness of the Yuzhnyi Chun Formation is up to 140 m. In the right-side area of the Ulakhan–Vava riverhead, the formation is lacking coarse tuffs, being composed of distal ashy tuffs and tuffites about 140 m thick in total.

Floral and faunal remains are scarce in the Yuzhnyi Chun Formation. In middle courses of the Sengachanda River (an isolated exposure on the right side), rocks representing an upper part of the unit yield bivalve species *Palaeanodonta biltchanica* Mal., *P. obrutschevi* (Rag.), *Abiella* (?) *elliptica* (Khalf.), and *A. subovata* (Jones). Ostracodes from this site represent *Darwinula decima* Misch., *D. fragilis* Schn., *D.* aff. *futchiki* Kasch., *D. inornata* (Spizh.), *D. inornata* (Spizh.) var. *macra* Lun., *D. undulata* Misch., *Suchonella cauta* Misch., *S. clivosa* Misch., *S. innominata* Misch., *S. stelmachovi* Spizh., *S. typical* Spizh., *Tatariella* aff. *wologodskiella* Misch., and *Gerdalia* sp. Associated conchostracans include *Rohdendorfium* (*Bipemphigus*) *tutontchanum* (Nov.), *Pseudestheria novacastrensis* (Mitch.)., *Glyptoasmussia belmontensis* (Mitch.), and *Trigonestheria permiana* Nov. Plant remains from the Sengachanda site are those of *Osmundopsis* ex gr. *kirjamkensis* (Pryn.) and *Cladophlebis* (?) cf. *grandifolia* Mog. Palynological assemblage from this site is of the Mesophytic appearance. Pollen distinctly prevailing over spores in this assemblage represent genera *Gnetaceapollenites, Vitreisporites*, and *Klausipollenites*. Dominant among spores are *Acanthotriletes* and *Osmudacidites* forms and *Lophotriletes tuberculatus* (Waltz). The Paleophytic palynological assemblage of *Raistrickia cornuta* (Andr.), *R. obtusosaetosa* (Lub.), *R. incurvispina* (Andr.), *R. grandispinosa* (Andr.), *Apiculatisporites multicolorius* (Andr.), *A. multiangulus* (Andr.), *Remysporites psilopterus* (Lub.), *Nigrisporites marginatus* (Portn.), *Calamospora* sp., *Zonomonoletes gracilis* Peters., *Circulisporites parvus* de Jersey, *Circella stenolimbata* Lub., *Subsacculifer incerches*

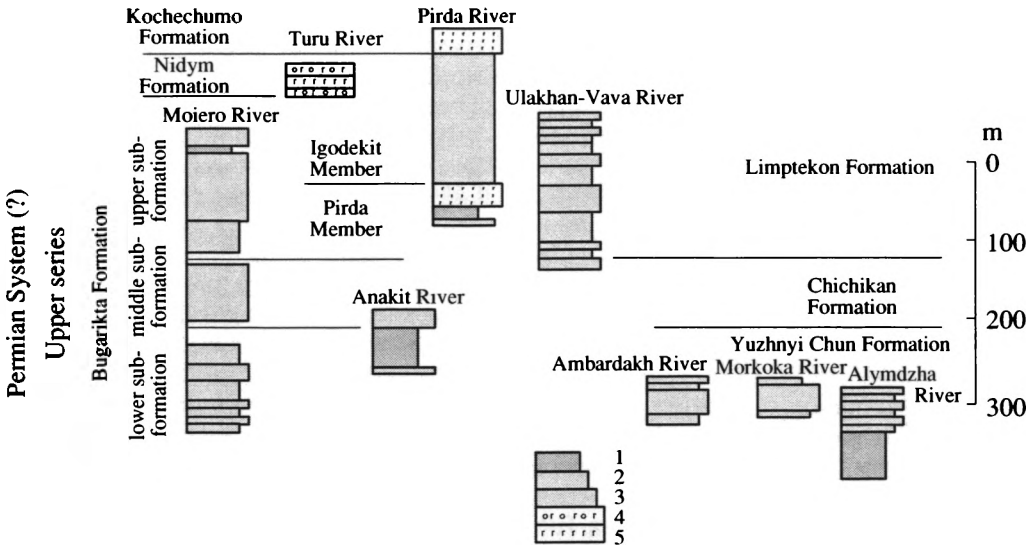

Fig. 6. Correlation scheme of principal Permian sections (above the Gagarii Ostrov Horizon): (1) ąshy tuffs and tuffaceous siltstones and shales; (2) tuffaceous psammite or sandstone; (3) coarse and finer tuffs of basic composition; (4) amygdaloidal lava; (5) basalt.

(Lub.), *Samoilovitchisaccites turboreticulatus* (Sam.), and of Cordaitales and *Ginkgocycadophytus* forms is known from several sites in the Ulakhan-Vava, Kyuryungnekyan, and Lakharchana river basins.

These paleontological and palynological data show that the Yuzhnyi Chun Formation is correlative with the Tutonchana or Log Lebedeva Horizon. Plant taxa listed above are comparable with those from the Log Lebedeva or younger stratigraphic levels. The formation can be undoubtedly correlated with the Vyatka Horizon of the upper Tatarian Substage.

In all the areas mentioned above, the Chichikan Formation either overlies (with a gradual transition) the Yuzhnyi Chun Formation, or overlaps older subdivisions of the Upper Paleozoic and even Silurian to Ordovician strata in the easternmost sites. Constituent rocks of the formation are medium- to coarse-grained tuffs usually lacking layered structure and intercalated with rare interbeds and lentils of tuffites and tuffaceous sediments. In the easterly sections (Alakit, Markha, and Ygyatta riverheads), they are 70 m thick. The formation thickness gradually grows southwestward: it is equal to 120 m in the Vilyui–Morkoka interfluve and ranges between 100 and 200 m (in places to 400 m) at the right side of the Vilyui River and further westward. Samples collected from several exposures of the Chichikan Formation yielded Paleophytic palynological assemblages similar to those of the Yuzhnyi Chun Formation.

In the eastern areas, both formations cannot be discriminated from each other based on data of geological survey, those of photointerpretation included, thus being shown in the map as undivided (*Geologicheskaya karta...*, 1991). In this undivided sequence, its lower part corresponding to the Yuzhnyi Chun Formation is composed of coarse tuffs in the Lakharchana riverhead and of layered tuffites and tuffaceous sediments in source areas of the Olguidakh and Ygyatta rivers. In other areas, coarse tuffs are typical of its middle member, whereas underlying and overlying beds are composed of fine-grained volcaniclastic sediments. Thickness of the undivided sequence ranges from 100 m, and less, to 200 m at a maximum.

In the Eika River basin and in the Turu and Ulakhan-Vava riverheads, rocks of the Limptekon Formation overlie either the Chichikan (usually), or the Yuzhnyi Chun (in places) formations. In the last two areas, the Limptekon Formation is composed of silty to psammitic volcanomictic sediments with subordinate interlayers of pelitic tuffites and tuffs. In other areas, its sequence also includes beds of coarse tuffs. In the Turu, Pirda, and Kananda riverheads, the lower (Pirda) member of this formation encloses some basaltic lava flows, and coarse tuffs are abundant in the upper (Igodekit) member. These members are 180 and 300 m thick, respectively.

In many exposures of the Eika River basin, rocks of the Limptekon Formation bear remains o fauna and flora. Bivalves from the Pirda Member are represented by *Palaeanodonta* aff. *bicarinata* Amal, *P.* aff. *aceris* Amal., *P.* aff. *subcastor* Amal., *P. subparallela* Amal, *Palaeomutela* ex gr. *rectodonta* Amal., *Anthraconaia vivi* Mal., and *Anthraconauta euri* Mal. They are associated with conchostracans *Rohdendorfium (Bipemphigus) tutontchanum* (Nov.), *Sphaerograpta dechaesauxae* Nov., *S. feysi* Nov., *Liograpta strictocostata* Nov., *Tripemphigus sibiricus* Nov., *Pseudestheria cicatricosa* Nov., *P. nordvikensis* Nov., *P. novacastrensis* (Mitch.)., *Megasitum petrovi* Nov., *Glyptoasmussia belmontensis* (Mitch.), *Limnadia* (*Falsisca*) *jeskinoica* Nov., *L.* (*F.*) *turaica* Nov., *L.* (*L.*) *glabra* (Mitch.), *L. monmogtchensis* Nov., *Taimyrites pruvosti* Nov., and

with ostracodes *Darwinula aemayica* Schl., *D. casta* Misch., *D. inornata* (Spizh.) var. *macra* Lun., *D. quadrata* Misch., *D.* ex gr. *temporalis* Misch., *D. tutontchaniensis* Misch., *Darwinuloides sentjakensis* (Schar.), *Suchonella cauta* Misch., *S. circulata* Misch., *S. clivosa* Misch., *S. stelmachovi* Spizh., *S.* ex gr. *subsimila* Misch., *S. subtilis* Misch., *S. typica* Spizh., *Tatariella citata* Misch., *T. permiensis* Misch., and *Gerdalia clara* Misch. Fish remains from the same site represent genera *Colobodus, Evenkia, Hybodus*, and *Tungusichtys*. In addition to this fauna, there were found very diverse plant remains: *Sphenophyllum thonii* Mahr f. *evenkense* Sadovn., *Equisetum* (?) *triassicus* (Radcz.), *Neokoretrophyllites linearis* (Pryn.), *Todites* cf. *crenata* (Pryn.), *T.* (?) ex gr. *fragilis* Daugh., *T.* cf. *lobifera* (Pryn.), *T.* ex gr. *whitbiensis* (Brongn.), *Acrostichides* cf. *kirjamkensis* (Pryn.), *A.* cf. *shvedovii* Sadovn., *A.* cf. *shensiensis* Sze, *A. srebrodolskae* (Shved.), *Mertensides* aff. *bullatus* Font., *M.* cf. *concinnus* Mog., *Discopteris* ex gr. *rotundiloba* (Font.), *Boweria* sp., *Sphenopteris* cf. *trisecta* Shved., *Pecopteris* ex gr. *neuburgiana* (Molot. et Tesl.), *Cladophlebis* ex gr. *augusta* Heer, *C. grabauiana* P'an, *C.* cf. *kaoiana* Sze, *C.* aff. *stenolopha* Brick, *C.* (?) ex gr. *undulata* Brick, *Katasiopteris* cf. *polymorpha* Mog., *Lepidopteris* sp., *Tatarina* sp., *Edyndella* cf. *dentate* Mog., *Glossophyllum* (?) sp., *Peltaspermum* sp., *Scytophyllum* cf. *tenuinervis* Mog., *Parajacutiella* (?) *parva* (Mog.), *P.* (?) *angusta* Mog., *Rhipidopsis* cf. *lobate* Heer, *Yuccites* (?) *angaridensis* Mog., *Carpolithes* ex gr. *cinctus* Nath., *C.* ex gr. *minor* Pryn., *Allicospermum* (?) sp., *Quadrocladus* cf. *pachyphyllum* (Pryn.), and *Q. sibiricum* (Neub.). Characteristic palynological assemblages are of the Mesophytic type. Rare Paleophytic assemblages from basal beds predominantly consist of *Vitreisporites pallidus* (Reis.) pollen associated with spores of *Turrisporites sibiricus* Medv., *Apiculatisporites spinosus* (Naum.), *A. globulosus* (Andr.), *A. rectispinus* (Lub.), *Cyclogranisporites larvatus* (Lub.), *C. polypyrenus* (Ibr.), and *Osmudacidites* species. Occurring sporadically are spores of *Nigrisporites nigritellus* (Lub.) and *Remysporites marginatus* (Portn.) and pollen of *Cordaitina praecipa* Medv., *C. rotata* (Lub.), *C. rugulifera* (Lub.), *Remysporites psilopterus* (Lub.), *Bennettites lautus* Medv., *Entylissa caperata* (Lub.), and *E. glabra* (Lub.). The Mesophytic assemblages consist of abundant palynomorphs representing taxa *Osmudacidites, Punctatisporites, Acanthosporites bradyensis* Playf., *Lophotriletes triassicus* Korotk., *Gnetaceapollenites, Vitreisporites koenigswaldii* Jans., *Lueckisporites*, and *Gardenasporites*.

The Igodekit member yields bivalves of genera *Palaeanodonta* and *Anthraconaia* and conchostracan taxa "*Rohdendorfium*" (*Bipemphigus*) *tutontchanum* (Nov.), *Hemicycloleaia, Kaltanleaia, Mimoleaia(?), Megasitum harmonicum* Nov., *Echinolimnadia mattoxi* Nov., *Limnadia (Falsisca) turaica* Nov., *L.* (L.) *glabra* (Mitch.), and *Cornia*. Concurrent ostracode forms are *Darwinula activa* Star., *D. angulata* Mand., *D. curvidorsalis* Mand., *D. lacrima* Star., *D mera* (Misch.), *D. minuta* Mand., *D. triassiana* Bel., *Gerdalia clara* Misch., *G. dactyla* Bel., *G. minuta* Star., *G. polenovi* Bel., *G. variabilis* Misch., and *Darwinuloides oviformis* Mand. Species *Stenochara elongata* Said., *S. maedleri* (H. et P.), *Porochara kiparisovae* Said., *P. lipatovae* Kis., *P. lutkevitchii* Said., *P. movschovitschii* Said., *P. rykovii* Said., *P. sphaerica* Kis., and *Cuneatochara acuminata* Said. represent charophytes of the unit, whereas remains of higher plants correspond to *Pleuromeia taimyrica* Sadovn., *Takhtajanodoxa* cf. *mirabilis* Snig., *Neocalamites* (?) cf. *triassica* Radcz., *Todites* (?) *fragilis* Daugh., *Acrostichides* (?) *kirjamkensis* (Pryn.), *A.* cf. *tchunicus* Sadovn., *Sphenopteris* ex gr. *trisecta* Schved., *Cladophlebis* ex gr. *haiburnensis* (L. et H.), *C.* ex gr. *ichiinensis* Sze, *C.* aff. *kirjamkensis* Pryn., *Lobatopteris multinervis* (Neub.), *Lepidopteris* sp., *Carpolithes* sp., *Quadrocladus sibiricum* (Neub.), and *Q.* aff. *pachyphyllum* (Pryn.). Paleophytic and Mesophytic palynological assemblages are similar to those of the Pirda Member.

In the Yunekeen River basin westward of the Pirda lavas distribution area, the Limptekon Formation bears ostracodes *Darwinula lucida* Mand., *Pyrevaria pyreniformis* Mand., and *Gerdalia clara* Misch. Associated plant remains represent *Boweria* sp., *Mertensides lingulates* Mog., *Boreopteris evenkensis* Mog., *Eleganopteris tripinnata* Mog. et Il., *Acrostichides srebrodolskae* (Shved.), *A.* (?) *tchunicus* Sadovn., *A. tunguskanus* (Pryn.), *Cladophlebis* cf. *kaoiana* Sze, *C.* (?) ex gr. *undulata* Brick, *Katasiopteris* cf. *lebedevii* Radcz., *Rhipidopsis lobata* Halle, *Quadrocladus* sp. Paleophytic and Mesophytic palynological assemblages are similar to those from the Pirda and Igodekit members. All these data imply that the Limptekon Formation can be attributed to the Khungtukun or Putorana horizons (Taimyr regional stage of the terminal Permian).

In the Southern structural zone, there are widespread lavas spanning approximately the range of the Nidym and Kochechumo formations. According to Cherenkov *et al.*, 1991), table mountains at the right side of the Vilyui River and in the Pasporin, Turu, and Ulakhan-Vava riverheads are composed of lavas at the top. Considering lavas and intercalated tuffs and tuffaceous sediments as rocks of the Anaun Formation (an analogue of the Bugarikta, Limptekon, or Nidym formations), they attributed these lavas to the facies of immense volcanic conduits. The eruptive origin of lava bodies has not been proved however. These bodies lacking amygdules are likely of intrusive origin, as one can judge from their persistent thickness, insignificant differentiation, and some other features.

Some researchers regarded the thick volcanic bodies as lava flows armoring from erosion the underlying rocks in the Vilyui–Morkoka interfluve (sites of the Syan River, Morkoka left bank, and Vilyui right bank) and in lower courses of the Ulakhan-Vava River. They argue for difference in physical properties of rocks con-

stituting these bodies and true sills intruding the upper Paleozoic sequence. I think that this viewpoint is misleading, because rocks of those bodies are considerably differentiated and Upper Paleozoic terrigenous to volcaniclastic rocks enclosing them are frequently transformed into hornfels of the outer contact zones.

The described volcanic rocks have been attributed to the Lower Triassic (*Resheniya tret'ego...*, 1981). This decision was based on the Mesophytic habit of fossil flora, and the rocks were correlated with Induan ammonite-bearing strata known in the eastern flank of the Orulgan Ridge. As is shown recently (Sadovnikov and Orlova, 1997), deposits in western slopes of the Orulgan and Kharaulakh ridges, which seem to be correlative with the easterly Induan strata (Nedzhela and partly Syurbelyakh formations), are undoubtedly younger than, but not concurrent with volcanogenic rocks of the Tunguska syneclise. Flora and fauna from lower beds of that volcanogenic sequence resembles to a great extent the late Tatarian floras and faunas of the Cis-Urals and Moscow syneclise. Being correlative with the upper Tatarian Substage, the sequence may also span a part of hiatus between the upper Tatarian and Lower Triassic strata. Its upper portion reveals alternation of "Permian" or "Triassic" assemblages of ostracodes and fossil insects, and conchostracan faunas from this level consist of either Permian, or endemic Siberian taxa. Consequently, this interval seems to be correlative either with a hiatus separating terminal Permian and basal Triassic beds, or with the Changsingian Stage of the Eastern Tethys. In general, volcanogenic rocks of the Tunguska syneclise are of the Permian age, although officially they are attributed to the Lower Triassic, being mapped in this rank (*Geologicheskaya karta...*, 1991).

Thus, coal-bearing Upper Paleozoic deposits in the eastern limb of Tunguska syneclise span the Middle–Upper Carboniferous and Permian intervals. Deposits of the first interval are much more widespread than the Permian sediments, among which the Lower Permian ones are of a very limited distribution (middle courses of the Vilyui River only). Lower beds of the Carboniferous interval, which are known under names Konek and Lapchan formations, should be jointly regarded under one of these names. The Term Aikhal Formation is more appropriate for the rest of the Carboniferous sequence than the Kata Formation established in the Angara River basin. The lower Botuoba Subformation corresponds in general to the Lower Permian, though its lower boundary should be corrected in those sections, where Carboniferous strata turned out to be placed above it. The upper Botuoba Subformation spans the terminal Lower and greater Upper Permian intervals. Abundance of coal seams and their metamorphic grade increase toward the axial syneclise zone, where coking coals may be of a wide distribution. Uppermost Paleozoic strata of predominantly volcanogenic rocks are incorrectly attributed to the Lower Triassic. These are tuffs and tuffaceous–sedimentary strata of the Korvuchana Group (three Bugarikta subformations in the west and Yuzhnyi Chun, Chichikan, and Limptekon formations in other areas). In the west, they are overlain by volcanogenic–sedimentary deposits and basaltic lavas of the Nidym and Kochechumo formations of the Putorana Group.

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Climate during the Permian–Triassic Biosphere Reorganizations. Article 2. Climate of the Late Permian and Early Triassic: General Inferences

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Abstract—The Late Permian–initial Triassic was a period of the Earth climate change, as the glacial climate of the Late Paleozoic was replaced by the non-glacial one of the Mesozoic. The intricate trend of this process is reconstructed and illustrated by schematic paleoclimatic maps. Warming in the second half of the Sakmarian Age resulted in a rapid degradation of huge glacial belt to the polar glacial cap. By the Kazanian Age, the latter was gradually replaced by the high-latitude temperate-cold belt with retained and new glacial centers that intermittently widened. In the Tatarian Age, a similar temperate-cold belt appeared in the Northern Hemisphere as well. The next strong and sudden global warming occurred in the Permian–Triassic boundary period when the temperate-warm climate developed in high latitudes of the Northern Hemisphere and temperate one in the Southern Hemisphere. General warming complicated by frequent different-rank oscillations transformed the climatic zoning on the Earth and resulted in rapid global ecological changes. Huge dimensions of Pangea and mountainous belts and ridges at its margins primordially determined wide development of the semiarid and arid climate in low latitudes. During the Late Permian–initial Triassic period, the aridity of the Earth was increasing that is evident from successive widening of arid and semiarid belts of Pangea, which advanced toward middle latitudes, and from grown aridity in the equatorial mountainous belt and in tropical latitudes of the Tethys. Global warming was main factor responsible for stepwise widening of arid and semiarid belts. Another cause of aridity increase was gradual regression of inner seas. Arid and semiarid belts occupied about 40% of land at the beginning of the Early Permian, 55% in the late Sakmarian–early Artinskian time, and 80% in the Induan Age. Paleoclimatic reconstructions confirm the assumed significance of monsoons impact on the climate development in the Permian and Early Triassic to a certain extent only. During glacial periods, the climatic asymmetry of the Earth was particularly remarkable. Registered environmental changes of the Permian time (regressions, aridity growth, land elevation, orogeny, and island-arc volcanism) could result only in cooling. Warming of terminal Permian–Early Triassic, as well as C, O, S, and Sr isotopic anomalies were probably the first results of the main Pfalzian phase of the Hercynian tectogenesis with associated weakening of suprasubduction volcanism, intense regional metamorphism, and denudation of carbonaceous sedimentary sequences of orogens. The subsequent outburst of mantle volcanism, in particular, trap eruptions in Siberia promoted these processes further. Depending on their scale, climatic changes prepared or even provoked biotic crises, the mass extinctions included.

Key words: Permian, Early Triassic, climate, climatic belts and zoning, glaciations, warming, aridity, oscillations.

INTRODUCTION

In our first article (Chumakov and Zharkov, 2002), we discussed purposes and methods of study and the Early Permian climate. This, second article is dedicated to the Late Permian and Triassic climate and to general inferences from the study. Similarly to the previous article, paleoclimatic reconstructions are based on stratigraphic correlation schemes proposed for the international project "Pangea" (Ross and Ross, 1995), maps of sedimentation environments (Zharkov and Chumakov, 2001), paleobiogeographic data (Dobruskina, 1982, 1994; Meyen, 1987; Leven, 1993; Durante, 1995; Wnuk, 1996; Grunt, 1995; Grunt and Shi, 1997; Yaroshenko, 1997; Rees *et al.*, 1999; Shi and Grunt, 2000; Ignat'ev and Naugol'nykh, 2001), and on other published paleoclimatic, paleobiogeographic, and geological materials. We should repeat also that unavoidable uncertainties in correlation imply that age limits of proposed paleoclimatic reconstructions are conditional to some extent, although they do not change their characterization and succession. In this article, similarly to the previous one, we pay main attention to climatic zoning because the latter reflects most adequately the planetary climate and energy and mass exchange in biosphere, i.e., its thermodynamic state. We selected for our paleoclimatic reconstructions two stratigraphic intervals, which best represent climatic zoning typical of the Late Permian and Early Triassic.

CLIMATIC ZONING IN THE LATE PERMIAN

At the beginning of the Late Permian, the climatic zoning resembled in general that of the terminal Early Permian, except for a slightly decreased asymmetry relative the equator and alike succession of belts in both hemispheres. Nevertheless, as is shown below, the width of corresponding belts, their latitudinal position, and relevant climatic parameters were notably different as before. At the beginning of the Late Permian, high latitudes in both hemispheres were occupied by temperate-cold belts with small-scale permanent or intermittent glaciation centers, whereas climate varied from the cold to temperate-cold. We begin characterization of climatic zoning with these belts gradually moving from high to low latitudes.

Northern temperate-cold belt of high latitudes. The belt is well distinguishable in the northern part of the northern humid belt (Zharkov and Chumakov, 2001) owing to a wide development of glacial and glaciomarine deposits of the Atka Formation in the Kolyma River upper courses (Epshtein, 1972; Chumakov, 1994). Similar deposits (diamictites) are known also in the western and southern parts of the Verkhoyansk fold belt (Dulgalakh Group, Andrianov, 1966, 1985; and others) and in the Omolon massif (Gizhiga Fomation, Kashik *et al.*, 1990). These units used to be considered as the late Kazanian in age, are now dated back to the early Tatarian according to present-day views substantiated mainly by paleomagnetic data (Kashik *et al.*, 1990; Kotlyar, 1997). In the Kolyma River upper courses and east of the Omolon massif, diamictites are also recorded in the overlying Khivach Horizon correlated with the upper part of the Tatarian Stage (Byakov, 2000). In addition to glacial and glaciomarine sediments, the relatively cold climate of the belt is evident from the impoverished state of concurrent Siberian vegetation (in Verkhoyansk okrug (district) of paleoflora after Durante, 1995, or in North Siberian region after Wnuk, 1996) and from occurrence of high-boreal invertebrate fauna (Grunt, 1995) and absence of conodonts (Wardlaw, 1995) in surrounding seas. Geological and paleontological data are well consistent with paleomagnetic and geodynamic reconstructions, according to which the belt in question was situated between 60° and 75° N (Kharamov *et al.*, 1982; Scotese and Langford, 1995; Parfenov *et al.*, 1999) or between 70° and 87° N (Ziegler *et al.*, 1998).[1] The Tatarian cooling responsible for accumulation of glacial and glaciomarine deposits was preceded probably by some warming of the Ufimian and Kazanian ages, because the Gizhiga Formation of the Omolon massif rests upon the Omolon Formation composed of limestones. In the fossil assemblage from the latter, prevalent are *Kolymia* forms of bivalves,[2] brachiopods, smaller foraminifers, bryozoans, and solitary rugose corals (Ganelin, 1984). These dominant groups indicate boreal (Ustritskii, 1993) or even high-boreal (Grunt, 1995) composition of the Omolon fauna. Some fossils from the Omolon Formation, such as *Atomodesma, Spiriferella, Uraloceras*, and other forms widespread in the southern temperate-cold belt (Shi and Grunt, 2000; and others), are of bipolar distribution. Accordingly, it is possible to assume that limestones of the Omolon Formation accumulated in moderately cool waters, like the Permian limestones of southeastern Australia and New Zealand (Runnegar, 1984), compositionally similar Kazanian deposits of the Sverdrup basin in Canada (Beauchamp and Theriault, 1994; Beauchamp, 1995), and the present-day Arctic biogenic carbonates of the Norwegian and Barents seas (Freiwald, 1998). Accumulation of calcareous sediments of the Omolon Formation could be also stimulated by a moderately warm current of the anticyclonic gyre similar to present-day North Pacific one, which is inferred for Panthalassa (Kiessling *et al.*, 1999) and whose branches could reach northwestern margins of Pangea during warming episodes. Small-scale warming events occurred episodically also in the second half of the Tatarian Age. This is evident from occurrence of rare coal seams in the upper part of the Dulgalakh Group of the western Verkhoyansk region (Ganelin, 1984).

Southern temperate-cold belt of high latitudes. This belt is established between 70°–75° and 90° S based on development of coal-bearing deposits in the Transantarctic Mountains (Retallack and Krull, 1999; Isbell *et al.*, 1997) and southeastern Australia (Retallack, 1999a), where they enclose abundant horizons of relatively mature humid soils and remains of the *Glossopteris* flora. Coal-bearing deposits of Australian lowlands formed in boggy forests composed mainly of *Glossopteris* forms, which are indicative of the prevailing temperate climate. Nevertheless, significant cooling episodes, which were frequent in this belt, resulted in formation of glaciers and seasonal ice. In southeastern Australia, signs of iceberg- and ice-rafted material (dropstones) are repeatedly recorded in the Ufimian, Kazanian (Crowell and Frakes, 1971; Dickins, 1996; Eyles *et al.*, 1997; and others), and Tatarian (Veevers *et al.*, 1994) sequences. At the very end of the Late Permian, climate here was sometimes close to the subarctic one. In opinion of Retallack (1999a), some coals formed in bogs similar to present-day shoestring bogs in tundra areas of northern West Siberia and lowlands surrounding the Hudson Bay. The Tatarian deposits are noted to enclose forest soils with permafrost signs and root remains of plants similar to *Gangamopteris* forms (Retallack, 1999b). According to Runnegar (1984), seas surrounding southeastern Australia in the Late Permian can be referred to the coldest-water Pacific province of the Notal realm.

In the Antarctic, the intermittently swamped lacustrine–alluvial plains of the Late Permian time were populated by *Glossopteris* and accumulated peat (Isbell and Cuneo, 1996). The occurrence of dropstones with

[1] Hereinafter, the position of belts is given in paleolatitudes.

[2] *Kolymia* may form biostromas called "Kolymia reefs" (Ganelin, 1997).

distinct signs of glacial abrasion in coal-barren interbeds indicate a short-distance transport of stones by seasonal ice (Smith *et al.*, 1998) or glacier ice fragments brought by melted glacier waters.

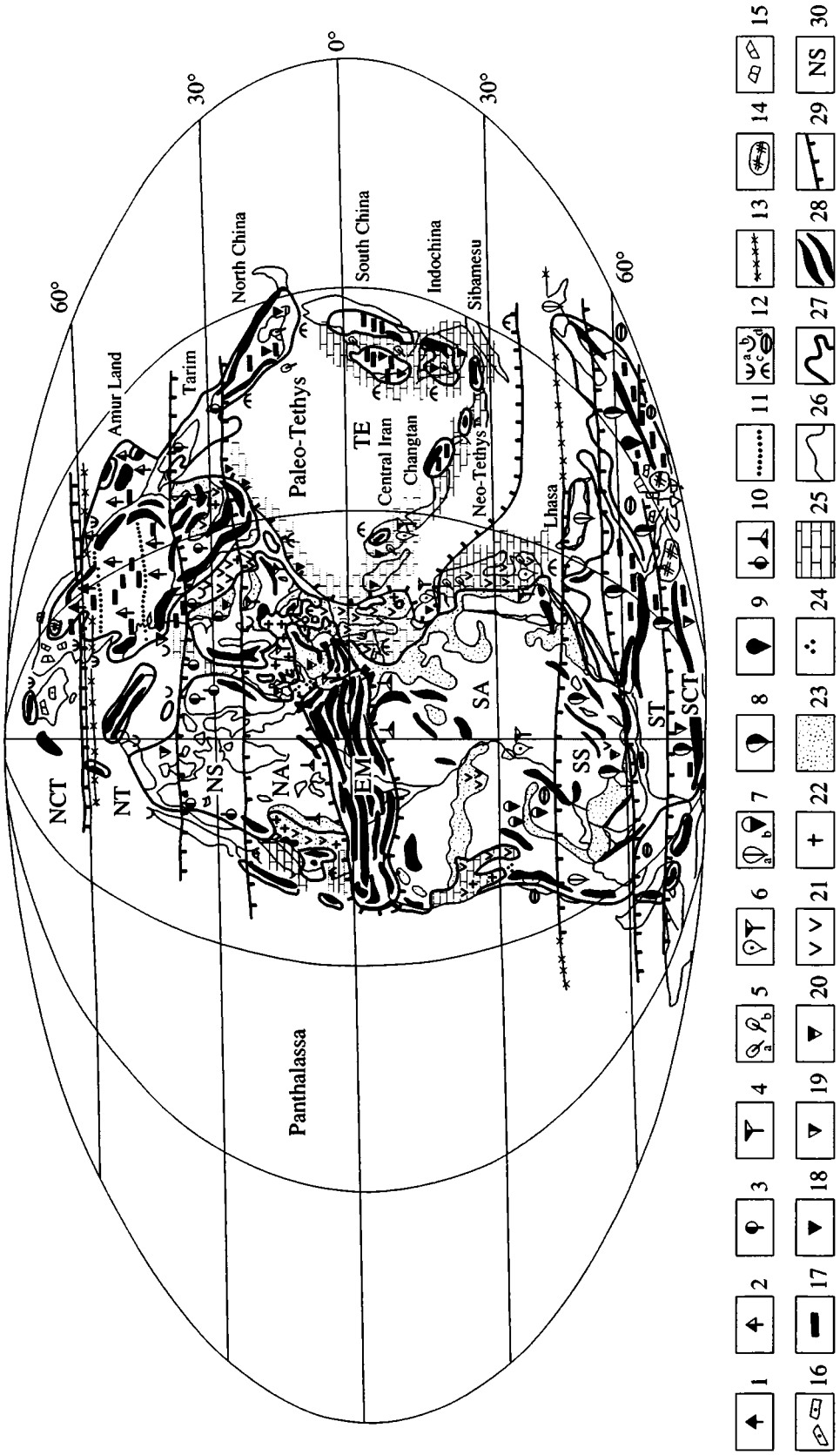
Data on cold-temperate climate in southern high latitudes are well consistent with abundant paleomagnetic data and paleogeographic reconstructions indicating that southeastern Australia and Transantarctic Mountains were located in the paleolatitude belt between 70° to 90° S (Scotese and Langford, 1995; Ziegler *et al.*, 1997).

Northern temperate humid belt of middle latitudes. In the Northern Hemisphere, the temperate humid belt was located approximately between 60° and 40° N (Fig. 1). In Asia, it occupied a southern part of the northern coal-bearing belt (Zharkov and Chumakov, 2001) and corresponded to the most southern part of the Siberian floral region, to its Tunguska okrug and Taimyr-Kuznetsk subregion. The latter included also the Amur Land, which likely joined Siberia in the east. Shelf seas of this belt were mainly populated by the low-boreal fauna (Grunt, 1995), which makes the northern temperate belt traceable westward to the Canadian Archipelago and northern Alaska. High-boreal assemblages of the shelf fauna are recorded in the Upper Permian deposits of the belt only near its northern boundary, in the Taimyr Peninsula area. The Kazanian deposits occurring in the marginal parts of the Sverdrup basin are represented by biogenic shelf limestones with the bryozoan–echinoderm–brachiopod assemblage. Based on lithological and paleontological peculiarities and on occurrence of dropstones in coastal terrigenous facies, these sediments are interpreted as deposits of moderately cold-water basins (Beauchamp and Theriault, 1994; Beauchamp, 1995). Rare dropstones in coastal sediments are likely indicative of episodic seasonal ice, formation of which, as is known, is possible in a typical temperate climate. The overlying deposits of the Sverdrup basin correlated with the Tatarian Stage are represented by both the deep- and shallow-water spicularites. This is interpreted as indicative of cold and even polar climate of the terminal Permian (Beauchamp, 1995). The interpretation is inconsistent however with occurrence of low-boreal invertebrate fauna in this region and with finds of Permian conodonts in northern Canada (Wardlaw, 1995) and the belt position between the temperate-cold and warm semiarid belts. Slight cooling proper in northern Canada accords well with some northward motion of Pangea during the Permian (Scotese and Langford, 1995; Ziegler *et al.*, 1997, 1998).

Southern temperate humid belt of high latitudes. In the Late Permian, the temperate-cold belt of the Southern Hemisphere bordered in the north with a narrow (15–20%) latitudinal belt located between 75 and 60–55° S, in which coal deposits and *Glossopteris* flora localities are abundant. Annual growth rings and layered accumulations of fallen leaves indicate the seasonal climate and presence of deciduous *Glossopteris* forests (Retallack, 1980). The shelf-sea fauna referred to the Indian Province of the Notal Realm was characterized by relatively thermophilic forms, some Tethyan elements included (Runnegar, 1984). Simultaneously, no signs of frosty climate are recorded within this belt, the climate of which can be classed with temperate and humid.

Northern warm semiarid belt. At the beginning of the Late Permian, this belt was located between 40 and 25° N being narrowed in its eastern part near the Paleo-Tethys coast (Fig. 1). Subsequently, the northern boundary of the belt migrated gradually northward, and the belt slightly widened. In the Eurasian segment, the belt was characterized by lacustrine–alluvial, variably calcareous terrigenous red beds and evaporite deposits, these with glauberite included, which imply alternation of arid cold and humid warm seasons (Zharkov and Chumakov, 2001). The deposits, particularly those correlative to the upper part of the Tatarian Stage, enclose red and variegated (red and gleyed) paleosols with illuvial horizons enriched in carbonates (Ignat'ev, 1962; Perel'man and Borisenko, 1999; Yakimenko *et al.*, 2000). Soils of this kind point to the semiarid climate with short seasonal and secular moistening periods. In the Tatarian Stage, there are rare interbeds of eolian sediments (Tverdokhlebov and Shminke, 1990). The terrestrial vegetation of the belt known as the Subangara flora and characterized by a substantial proportion of coniferous and thermophilic forms is also indicative of prevalent arid and sufficiently warm climate (Meyen, 1987; Durante, 1995). Some authors believe that climate in that belt was of the Mediterranean type (Ziegler, 1990). Finds of the Angara-type floral remains, which are of the Kazanian age in northern Greenland (Wagner *et al.*, 1999) and of the Ufimian age in the Canadian Arctic Archipelago (Roadian, LaPage *et al.*, 1999), and concurrent palynomorphs (Utting and Piasecki, 1995) make this belt traceable in the northern part of North America (Fig. 1). Inner seas of the belt with carbonate and evaporite platforms were populated by low-boreal invertebrate fauna (Grunt, 1995).

Southern temperate semiarid belt of middle latitudes. This belt was located between 60 and 45° S in western Gondwana and between 55 and 45° S in its eastern part, being slightly narrowed in the area of the Neo-Tethys southern coast because of the ocean influence as we suspect. The semiarid belt was most distinct in South Africa, where fans, alluvium, and deltas of intermittent water flows formed in piedmonts of rises and alluvium of meandering rivers, playa, and lacustrine deposits accumulated in plains (Zharkov and Chumakov, 2001). Lacustrine sequences host sometimes the black-shale members and carbonate interbeds, whereas alluvial deposits contain gouge, thin coal seams, and siderite interlayers. Also common are red to variegated rocks and thick (up to 4 m) carbonate paleosols with fossil root systems, rhizoconcretions, and small therapsids buried in hollows (Smith, 1990). Some


60°
30°
0°
30°
60°
Amur Land
Tarim
North China
South China
Indochina
Sibamesu
Paleo-Tethys
TE
Central Iran
Changtan
Neo-Tethys
Lhasa
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paleosols resemble caliche, while others are distinctly related to playas and exhibit desiccation cracks, well-developed clayey stress cutans (slickensides), and gypsum rosettes.

The belt in question was characterized by the *Glossopteris* and Australia–Africa–American (Wnuk, 1996) moderately thermophilic vegetation with abundant tree-ferns and lycopods. The latter flora recorded in Argentine, Tibet, and New Guinea likely points to a slightly higher moistening in the areas located closer to the ocean as compared with the glossopterid flora of inner areas of South Africa. Shelf seas of the belt were occupied by invertebrate fauna of the Notal type (Grunt, 1995). The absence of conodonts remains (Wardlaw, 1995) indicates probably the influence of cold currents.

Northern hot arid belt of low latitudes. The belt occupied areas from the piedmonts of mountainous Central Pangea (i.e., from 0–15° N) to northern tropical paleolatitudes (22–25° N). The land segment of the belt accumulated red alluvial–lacustrine deposits (Zharkov and Chumakov, 2001) enclosing abundant carbonate variegated to red paleosols (Mader, 1992) and, locally, gypsum-bearing sabkha and eolian sediments. Evaporite–carbonate platforms formed in inner and marginal parts of shelf sea basins (Fig. 1). The well-known Zechstein basin of Europe was most spacious saliferous basin within this belt at that time. The land segment of the belt was populated by the Euroamerian flora, and the Tethyan fauna inhabited seas. The low-boreal marine fauna dwelt only near the northern boundary of the belt throughout the entire Late Permian period (Grunt, 1995).

Sedimentation patterns, vegetation, and faunas of the northern arid belt imply that its climate was relatively hot, and this is consistent with the belt position in low latitudes.

Southern arid belt of low and middle latitudes. The northern arid belt extended along the eastern piedmont area of Central Pangea to equator, where it joined a spacious southern arid belt that occupied the large northern part of Western Gondwana. The latter extended from the western to eastern coast of the supercontinent between the piedmont of mountainous Central Pangea (0–10° S) and 45° S. Sedimentation in this belt was similar in general to that of the northern arid belt: prevalent were red lacustrine–alluvial sediments, frequently gypsum-bearing sabkha-related and eolian. Permian macro- and microfloras were scarce within the belt, likely confined to oases and valleys of transitory rivers. Vegetation was similar to the Euramerian flora in the north and east, and to the Gondwanan or mixed Euramerian–Gondwanan one in the south (Broutin *et al.*, 1990). Judging from its appearance, the Late Permian Euramerian flora characterizes the hot arid climate and, probably, with humid seasons in some areas (Wnuk, 1996). South America was populated by the aforementioned Argentine–Brazilian flora also of the desert type (Wnuk, 1996). Elements of the hygrophilous Cathaysian flora occurred only in the mixed Gondwanan–Euramerian vegetation of the Arabian Peninsula, in coastal areas of Pangea in the extreme southeastern part of the belt.

Shelf and inner seas of the arid belt hosted carbonate and carbonate–evaporite platforms. Waters along the entire Pangea eastern coast were populated by the characteristic Tethyan invertebrate fauna (Grunt, 1995). Notal forms prevailed only along its southwestern coast in South American seas near the southern boundary of the belt.

Substantial paleophytogeographic and paleozoogeographic differences within the belt of a large width allow an assumption that climate was non-uniform with respect to moisture and temperature. Climatic conditions ranged probably from temperate and more arid on the southwest to hotter and more humid on the east along the coast. Such a difference could be determined by anticyclonic oceanic currents: the cold southern near South America, i.e., along the eastern Panthalassa

Fig. 1. Climatic zoning of the late Kazanian-Tatarian time and paleontological of lithological indicators of climate: (1) moderately cold-resistant vegetation of the Verkhoyansk district of the Siberian phytogeographic region (after Meyen, 1987; Durante, 1995; and others); (2) moderately thermophilic vegetation of the Siberian phytogeographic region; (3) thermophilic vegetation of the Subangara phytogeographic region; (4) thermophilic xerophilous vegetation of the Euramerian phytogeographic region and alike vegetation of the Arabian Peninsula (after Wnuk, 1996; and others); (5) thermophilic and hygrophilous vegetation in the North (a) and South (b) China regions of the Cathaysian phytogeographic kingdom (after Wnuk, 1996; and others); (6) thermophilic vegetation of the Southern Hemisphere (Gondwanan vegetation with Euramerian and/or Cathaysian elements in Africa); (7) moderately thermophilic vegetation in the Australia-Africa-American (a) and Argentine-Brazilian (b) regions of the Gondwanan phytogeographic kingdom (after Wnuk, 1996); (8) vegetation of the temperate climate in the Gondwanan phytogeographic kingdom (*Glossopteris* region after Retallack, 1980, 1999; Wnuk, 1996; and others); (9) cold-resistant vegetation of the Gondwanan phytogeographic kingdom (*Gangamopteris* region after Retallack, 1980, 1999, Wnuk, 1996; and others); (10) main localities of palynoflora in indicated phytochores (after Utting and Piasecki, 1995; and others); (11) boundaries of some phytochores; (12) high-boreal (a), low-boreal (b), tropical/equatorial (c), and Notal (d) assemblages of shelf invertebrate fauna (after Grunt, 1995; Grunt and Shi, 1997); (13) northern and southern distribution boundaries of Early Permian conodonts (after Wardlaw, 1995); (14) assumed glaciation centers; (15) mostly marine and glaciomarine sediments; (16) the same, assumed; (17) coals; (18) bauxites or laterites; (19) humid soils; (20) carbonate, red-colored or variegated (partly gleyed) soils and calcretes; (21) gypsum and anhydrite; (22) halite, potassium and other salts; (23) arid calcareous red beds; (24) eolian sediments; (25) carbonate platforms; (26) present-day shoreline segments; (27) ancient shorelines; (28) mountainous systems and ridges; (29) boundaries of climatic belts; (30) northern and southern temperate-cold (NCT, SCT), northern and southern temperate (NT, ST), northern and southern semiarid (NS, SS), northern and southern arid (NA, SA), equatorial-mountainous (EM); tropical-equatorial monsoon-type (TE), and equatorial humid (EH) climatic belts.

coast, and the warm northern near the Africa–Arabia along the western Tethyan coast.

Belt with the equatorial mountainous climate. By the Late Permian, the mountainous area of Central Pangea was displaced slightly northward according to paleomagnetic evidence. Extending in the east-northeast direction, it crossed equator and was located in near-equatorial latitudes between 5–10 and 12–15° N. The piedmont and intermontane areas of the belt continued accumulating red beds with frequent signs of eolian reworking, horizons of carbonate, partly gleyed soils, caliche, gypsum, tetrapod traces (Mader, 1992; Cassinis *et al.*, 1995), and with rare plant remains of the arid Euramerian flora. It can be therefore assumed that mountain rivers mostly dried up at piedmonts even in case of intense atmospheric precipitation in mountains of Central Pangea, as it follows from mathematical modeling of the Late Permian climate (Kutzbach and Ziegler, 1993; Barron and Fawcett, 1995; Rees *et al.*, 1999).

Tropical-equatorial hot humid belt. The belt included the Paleo-Tethys that narrowed in the Late Permian, a greater part of the Neo-Tethys that widened at the expense of the latter, islands separating and fringing both oceans, and also their northwestern and, partly, southwestern coasts. As compared with the Early Permian, the belt widened slightly southward because of the warm anticyclonic current that originated in the Neo-Tethys. Thus, the belt partly occupied middle southern latitudes in its widest eastern part (Fig. 1). Continental sediments of the belt bear numerous signs of hot and humid climate: bauxites, remains of diverse Cathaysian vegetation (Wnuk, 1996), and coals. In southern China, the primary coal material accumulated mostly in tidal zones of carbonate platform and mangrove bogs (Shao *et al.*, 1998). Carbonate platforms were widespread in seas adjacent to Pangea and around microcontinents inside the Tethys. Nearly throughout the Late Permian, seas within the belt were populated by diverse thermophilic (Tethyan) fauna (Leven, 1993; Grunt, 1995, Grunt and Shi, 1997). In response to a global cooling in the terminal Late Permian, boreal forms became invading the Tethys to reduce diversity of the thermophilic fauna, and at the end of the epoch the reefal faunal community was preserved only in the eastern Tethys and adjacent margins of Panthalassa (Kozur, 1998).

CLIMATIC ZONING IN THE INITIAL TRIASSIC

The Permian–Triassic boundary period was marked by strong global warming, as it is evident from cessation of glaciation in polar areas and from thermophilic flora migration toward the high latitudes. The climate change was very rapid and led to substantial reorganization of the climatic zoning, new patterns of which indicate that the glacial climate of the Earth gave way to the non-glacial one, and that aridity in Pangea increased further. The zoning became so much symmetrical (Fig. 2) that it is possible to consider climatic belts of the Northern and Southern hemispheres together.

Northern and southern humid belts of high latitudes. Warming was particularly strong in high latitudes of the Northern Hemisphere. In Siberia of northern Laurasia, the Permian cordaitalean flora characterizing the temperate-cold climate was replaced by the Early Triassic coniferous–fern flora that points to expansion of ferns of the Cathaysian tropical and equatorial origin northward to the Taimyr and Verkhoyansk regions (Dobruskina, 1982), i.e., to 70–75° N. The palynologic assemblages from these regions are dominated by spores of hygrophilous plants (Yaroshenko, 1997). Thus, it can be stated that the climate of North Siberia became at least temperate-warm (Ziegler *et al.*, 1993, or even "tropical" (Dobruskina, 1994), and mainly humid at the beginning of the Triassic (Fig. 2, NWT). Simultaneously, the phytogeographic differentiation in Eurasia substantially decreased in general, which resulted in lower ranks of defined phytochores. Instead of three Permian floral kingdoms, the single Laurasian Kingdom with two the Angara and Euramerian floral regions is defined based on leaf remains in North and Central Eurasia for the Early Triassic (Dobruskina, 1982)[3] though three the Angara, Subangara, and Euramerian regions are defined based on palynocomplexes (Yaroshenko, 1997). The belt with the temperate-warm climate occupied the northern part of the Angara region. The northward advance of thermophilic flora and weakened floral differentiation reflects a substantial decrease of the latitudinal climatic gradient in Laurasia and the more gradual transitions between all climatic belts.

In southeastern Australia and Antarctica, the Early Triassic warming was responsible for replacement of the temperate-cold climate by the "temperate-cool" one, as it is evident from the changed composition of vegetation and paleosols (Retallack, 1999b; Retallack and Krull, 1999). Rare frost-related deformations in annual growth rings of fossilized trunks of Early Triassic trees (Jefferson and Taylor, 1983) suggest that negative winter temperatures were untypical of southernmost areas of Antarctica, which seems surprising because these areas were located at very high paleolatitudes (80–85°). Taking into consideration this and the other fact that during the preceding Tatarian Age the Antarctica continent hosted glaciation centers, which provided material for the ice rafting, one can believe that the warming amplitude in southern high latitudes was substantially high. According to some estimates (Retallack, 1999b), the temperature growth was as high as 6–11° in southeastern Australia, which might correspond to the continent migration for 15–20° of the latitude.

[3] Renamed subsequently into the Siberian and Eurosinian regions (Dobruskina, 1994).

The peculiar feature of both temperate belts of the Early Triassic Earth was a complete cease of coal formation. In the northern temperate belt, this could be, at least partly, a result of intermittent origination of arid settings, as one can judge from occurrence of red-bed members (Sadovnikov and Orlova, 1997; and others) and from admixture of pollen of xerophytes (Yaroshenko, 1997). The southern temperate belt was probably somewhat more humid that is supported by occurrence of coaly siltstones in the lower part of the Triassic section and by appearance of red beds only since the Middle Triassic (McLoughlin *et al.*, 1997; Retallack and Krull, 1999).

Northern and southern warm semiarid belts of middle and high latitudes. In the Early Triassic, the semiarid climate extended to middle and, partly, to high latitudes (Fig. 2, NS and SS). As a result, both these belts of the Earth widened significantly to become as wide as 40° of latitude. Like in the Late Permian epoch, the belts were distinguished by distribution of characteristic red-colored, locally gypsum-bearing, lacustrine, and alluvial deposits, among which a substantial proportion was represented by sediments of intermittent and wandering flows and of lakes with changeable shorelines and seasonal (Zharkov and Chumakov, 2001). These deposits frequently enclose calcareous red-bed and partly gleyed variegated buried soils (Chalyshev, 1968; Smith, 1990; Beauchamp, 1995; Tverdokhlebov, 1996). In the Moscow syneclise, the latter are sometimes represented by palygorskite varieties (Blom, 1972), which are relatively widespread now in semidesert regions. In the South Urals, there are signs of substantial high-mountainous glaciation in the Hercynian mountains (Tverdokhlebov, 1971).

The large width of semiarid belts determined their non-uniform climate and intricate phytogeographic patterns. This is particularly well exemplified by the northern semiarid belt. The Angara temperate-thermophilic and relatively hygrophilous flora was characteristic of its northern part, the mixed thermophilic to xerophilous flora of Angara and Subangara regions populated the central part, and the mixed flora of Subangara and Euramerian tropical arid areas, which appeared there since the Olenekian Age, occupied the southernmost zone (Yaroshenko, 1997; Dobruskina, 1982). The relatively gradual transition between phytochores and climatic belts confirms a widely accepted opinion that the latitudinal temperature gradient was insignificant at the beginning of the Triassic (Vakhrameev, 1985). This is consistent with uniformity of the Induan marine invertebrate fauna, which was almost similar in composition over the entire territory from the Verkhoyansk region and Greenland to the Himalayas (Dagys, 1976; Nevesskaya, 1999; Kozur, 1998).

Northern and southern arid belts of low latitudes. In the Induan Age, these belts widened slightly as well, although to a lesser extent than semiarid belts (Fig. 2, NA, SA). The arid climate was characteristic in general of low latitudes in both hemispheres, except for mountainous Central Pangea, the humid zone of the Tethys, which narrowed substantially in response to aridity increase in many coastal areas, and microcontinents of the North and, partly, South China. Arid belts were characterized by wide distribution of red beds and variegated, frequently eolian (with fossil dunes) and gypsum-bearing sabkha deposits in desert areas of land (Mader, 1992; Zharkov and Chumakov, 2001), whereas development of evaporite-carbonate gypsum-bearing platforms was characteristic of adjacent shelf seas. Continental sequences frequently enclose carbonate paleosols and calcretes, red-colored and variegated, with gleying structures (Mader, 1992). A remarkable feature of arid and semiarid belts of the Induan Age was absence of saliferous basins. It is difficult to explain this phenomenon taking into consideration the substantial enlargement of arid areas.

Because of scarce plant remains, vegetation of the belts in question is poorly known, particularly that of the very beginning of the Triassic. The Induan palynologic assemblages of the Euramerian affinity are found in some areas of the northern arid belt (Yaroshenko, 1997). Fossil flora of the Olenekian Age is known better (Dobruskina, 1994; Yaroshenko, 1997). This type flora characterizing the Early Triassic vegetation of the Euramerian region includes (in addition to abundant remains of cosmopolitan *Pleuromeya*) xerophilous coniferous and some other plants of the *"Voltzia"* assemblage, being very similar to the Upper Permian Zechstein flora. The similarity implies hot and arid climatic conditions in the belts during the Induan Age (Dobruskina, 1994). It is believed (Vakhrameev, 1985) that the Euramerian vegetation did not form a continuous cover and grew mainly in oases and along seacoasts.

Belt of the equatorial mountainous climate. Except for a slight northward migration, this belt remained without noticeable paleogeographic and climatic changes in the Early Triassic. As before, it was characterized by an intricate vertical climatic zoning with arid piedmonts.

Equatorial humid belt. As was mentioned, the vast Permian tropical-equatorial belt with the humid climate in the Tethys became substantially reduced because of the aridity increase in northwestern coastal areas of the Paleo-Tethys, northern microcontinents of the Cathaysian arc, and along western coasts of the Neo-Tethys. The area with the humid climate was transformed into a narrow near-equatorial belt as wide as 15° of latitude only (Fig. 2). The climate humidity is evidenced by fossil plants from the Lower Permian deposits of the Khainan Island, which undoubtedly represent descendants of the Cathaysian *Gigantoperis* flora characterizing tropical forests, and by bauxites in Central Iran and Turkey.

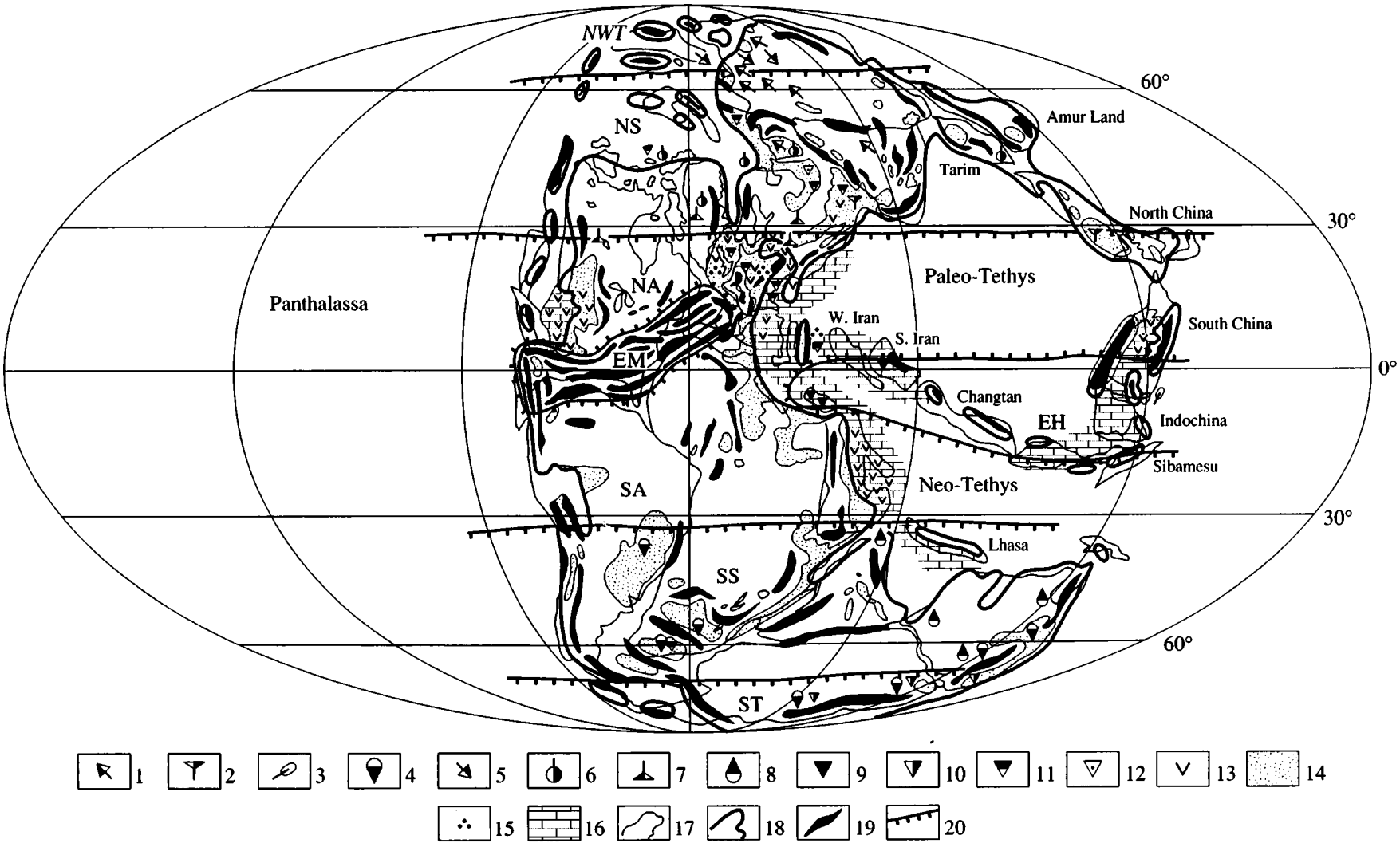

Fig. 2. Climatic zoning of the Induan Age and paleontological or lithological climatic indicators: (1) flora of the Angara phytogeographic region (Dobruskina 1982, 1994); (2) flora of the Euramerian phytogeographic region (Dobruskina 1982, 1994); (3) flora of the Cathaysian origin (Ziegler *et al.*, 1993); (4) flora of the Gondwanan phytogeographic kingdom; (5) palynoflora of the Angara region (hereinafter in terminology of Yaroshenko, 1997); (6) palynoflora of the Subangara region; (7) palynoflora of the Euramerian region; (8) palynoflora of the Gondwanan phytogeographic kingdom; (9) bauxite or laterite; (10) humid, (11) calcareous red to variegated (partly gleyed), and (12) palygorskite soils or calcretes; (13) gypsum and anhydrite; (14) arid calcareous red beds; (15) eolian sediments; (16) carbonate platforms; (17) present-day shoreline segments; (18) ancient shorelines; (19) mountainous systems and ridges; (20) boundaries of climatic belts. Climatic belts: (NWT) northern temperate-warm, (ST) southern temperate, (NS, SS) northern and southern semiarid, (NA, SA) northern and southern arid, (EM) equatorial-mountainous, (TE) tropical-equatorial of monsoon type.

GENERAL INFERENCES

The materials discussed in this and previous article suggest the following inferences on climatic changes that occurred in the Permian and initial Triassic epochs and on their possible causes and some consequences.

Warming. The main tendency during the geological period under consideration was warming, which so much changed climatic conditions on the planet that the glacial climate of the Late Paleozoic was replaced by the Mesozoic non-glacial climate. The discussed data offer a possibility to trace development of this intricate process in detail. During the glacial maximum of the Asselian and initial Sakmarian ages, spacious glaciation spread over the high and, to a considerable extent, in the middle latitudes of Gondwana and over southern microcontinents of the Cimmerian arc. After some oscillations, glaciers retreated to high latitudes in response to strong warming in the second half of the Sakmarian Age. Subsequently, at the beginning of the Artinskian Age, the retreat of glaciers slowed down. At that time, they occupied mainly the southern trans-Polar region. Thus, the glacial sheet was transformed into a polar ice cap that widened episodically advancing into lower latitudes, where local glaciation centers appeared sometimes (Visser, 1997). The further degradation of Gondwana glaciers occurred probably at the end of the Early Permian. Nevertheless, the glacial sheet remained in Antarctica at that time and related glaciers experienced repeated oscillations as can be inferred from the regular transport of iceberg material into southeastern Australia (Eyles *et al.*, 1997). Two last active phases of Gondwanan glaciers occurred in the Kazanian and Tatarian ages, respectively. The first of them is recorded in southeastern Australia by frequent iceberg- and seasonal ice-rafted materials, by activation of mountainous glaciation (Crowell and Frakes, 1971; Veevers *et al.*, 1994; Crowell, 1999; and others), and, at the very end of the Permian, by development of permafrost (Retallack, 1999a). This cooling was probably of global scale and relatively strong, because glaciomarine sediments (Epshtein, 1972; Chumakov, 1994) became widespread in high latitudes of the Northern Hemisphere approximately at the same time, in the Tatarian Age. In addition, the global scale of the latest Permian cooling is evident from migration of a high-latitude marine fauna to middle and low latitudes, and from extinction and distribution area reduction of some thermophilic forms in the Tethys (Kozur, 1998).

The second (after the middle Sakmarian), very sharp and strong warming that occurred in the Permian–Triassic boundary period resulted in complete degradation of planetary glaciation, and the temperate-warm (Ziegler *et al.*, 1993) or even probably "tropical" (Dobruskina, 1994) climate spread during a very short time over high latitudes of the Northern Hemisphere and the temperate climate became established in high latitudes of the Southern Hemisphere (Retallack, 1999b). This implies the stepwise increase in average annual temperatures at these latitudes by 8–15°C. Even in low latitudes, the temperature rise could be as high as 5°C in the Permian–Triassic boundary interval (Holster *et al.*, 1989).

Aridity increase. Huge dimensions of Pangea and spacious Hercynian mountainous belts and ridges, some of which were located at the supercontinent margins, sheltered the intracontinental areas from the oceanic moisture transport and determined a priori the wide development of semiarid and arid conditions in low latitudes of supercontinent. The geological period under consideration was a time of progressive aridity growth in Pangea, as it is repeatedly noted (Robinson, 1973; Parrish, 1995; and others) and confirmed by available reconstructions. This process led to successive widening of arid and, in particular, of semiarid belts, which advanced to middle latitudes, and to increasing aridity in the equatorial mountainous region of Central Pangea and tropical latitudes of the humid Tethyan realm (Fig. 3). Strong global warming and related intense evaporation represent an obvious factor of climate aridity. This is evident from the fact that both episodes of significant warming (the mid-Sakmarian and Permian–Triassic boundary periods) were accompanied by stepwise widening of arid and semiarid belts of Pangea (Fig. 3). After the second one, arid climate spread also over the Tethyan segment of the Earth (Figs. 1 and 2). The arid and semiarid belts occupied about 40% of land at the beginning of the Early Permian, while that proportion increased to 55 and 80% after the late Sakmarian and Permian–Triassic warming episodes, respectively. Successive regression of Pangea inner seas and associated disappearance of intracontinental moisture sources could represent another possible cause responsible for the emergence of arid climate. According to some researchers, the aridity of Pangea was mainly caused by monsoon circulation over the formed supercontinent that gradually moved northward (Robinson, 1973; Parrish, 1995). Paleoclimatic reconstructions confirm this view only in part. The monsoon influence was perceptible along the Tethys coasts, whereas Laurasia was probably sheltered from monsoons by young marginal ridges and Gondwana was protected by high pressure in the atmosphere above the ice shield and, later on, above tropical areas. Signs of monsoons influence have not been recorded even in the Permian and Early Triassic sequences of intermontane depressions and southern piedmonts of high-mountainous Central Pangea, where monsoons should discharge the main share of moisture according to available views on and mathematical modeling of the Permian climate (Kutzbach and Ziegler, 1993; Barron and Fawcett, 1995). Only at the very beginning of the Permian (Asselian and early Sakmarian time, Autunian, Wolfcampian), when sea gulfs still existed in piedmonts of high mountains, the humid climate could intermittently develop in Central Pangea.

Climatic asymmetry. A slight climatic asymmetry relative to equator is characteristic now of the Earth and some other planets. It is probably related to asymmetrical distribution of continents and oceans or, in a broad sense, to the asymmetrical orography of planets. One can assume therefore that the climatic asymmetry of the Earth existed always and approached an extreme state during glacial periods. This is particularly well exemplified by the Permian–Triassic climatic history. The Early Permian glaciations developed in high and middle latitudes of southern continents and, locally, almost reached the southern arid zone. The Northern Hemisphere was probably free of glaciers, and even if they existed, as some researchers assume, the glaciation was virtually negligible. Provided a similar or unchanged distribution of continents and orographic structures, the climatic symmetry became restored immediately after complete degradation of glaciations in the terminal Permian (Figs. 2, 3). This indicates two important properties of climatic system of our planet: first, the strong climatic asymmetry can appear on the glacial Earth only; and second, climatic systems of the Northern and Southern hemispheres are autonomous to a considerable extent.

Reorganization of the climatic zoning. Reduction of glacial sheets and subsequent cessation of glacial regime, warming, aridity increase, and restoration of climatic symmetry changed substantially climatic zoning of the Earth. Reorganizations in Pangea were particularly significant. Every large warming changed latitudinal position, width, main parameters, and number of climatic belts (Fig. 3). Most significant changes occurred within high and middle latitudes of the supercontinent. The low latitudes were mainly marked by changes in the width of belts. Climate was most stable in the Tethyan segment of the Earth, where the humid, sufficiently warm tropical and equatorial climate prevailed nearly throughout the entire Permian epoch. Nevertheless, at the Permian–Triassic boundary time, areas of humid climate became reduced even there, and humid conditions remained only in the narrow near-equatorial zone (Fig. 2).

Evolution of global climate. As climatic zoning depends on the planetary climate, its distinct transformation recorded across the Permian–Triassic boundary point to the drastic change of climate on the globe. The Permian climate should be classed with the glacial one, because glaciers and cold climate in high latitudes existed until the end of this epoch. This also means that psychrosphere in the ocean also existed until the end of the Permian. Its existence until the Kazanian Age at least is evident from the occurrence of relatively cold-resistant boreal conodonts of the Wordian Age in bottom waters of the tropical zone (Kozur, 1998). Therefore, the psychrosphere could exist during the subsequent cooling at the end of the Permian as well. Three types of the global glacial climate can be distinguished based on the character of Permian glaciations. The Asselian-Sakmarian climatic environments can be termed as *climate of the glacial maximum.* In popular publications, the glacial maximums are frequently termed as *great glaciations.* Environments of the terminal Sakmarian-initial Artinskian time, when glaciation was restricted to high latitudes can be termed as the *climate of ice caps.* It was similar to the present-day or to the Oligocene and late Eocene climate, when first ice shields appeared in Antarctica. Beginning from the second half of the Early Permian, the temperate-cold climate became prevalent in polar areas, although small glaciation centers (probably episodic) continued to exist in Antarctica and even new glaciation centers appeared in northern Laurasia, for instance, in the Kolyma River basin. This time can be considered as a final stage of the Gondwanan Glacioera with the *climate of temperate-cold polar areas.* The *non-glacial arid climate* of the Early Triassic, as one can judge from its main parameters, initiated the long-lasted Siberian thermoera, as it has been termed (Chumakov, 1984).

Dynamics of climatic changes. The discussed general trend of climatic changes during the Permian and Triassic was not uniform and unidirectional. In addition to main episodes of very rapid warming in the mid-Sakmarian and at the Permian–Triassic boundary time, the trend was complicated by numerous oscillations of different ranks. The high-rank oscillations are most distinctly recorded in the Permian glacial sections, where events that lasted several dozens of millions, several millions, and hundreds of thousands years are recognizable (Chumakov, 1985; Dickins, 1985; Veevers and Powell, 1987). The formation periods of Upper Paleozoic cyclothems in North America, which reflect, in opinion of many researchers, eustatic and climatic fluctuations related to oscillations of Gondwanan glaciers, also lasted several hundreds of thousand years (Veevers and Powell, 1987). Lower-scale climatic fluctuations are registered in some sections of ice-free areas. For instance, the Castile Formation of the Ochoan Group (the upper Tatarian Substage) in North America shows the rhythms that are believed to correspond to climatic oscillations of 200, 100, 20, and 2.7 kyr (Anderson and Dean, 1995). Most of these oscillations are close to orbital Milankowitch periods.

Oscillations imparted even a greater dynamics to rapid climatic changes of the Permian epochs, particularly during the glacial periods. The high formation and even higher degradation rates of ice covers are well exemplified by Pleistocene glaciations, which resulted in rapid ecological changes of subglobal scale. Similar processes occurred probably during the Permian glacial periods, as it is evident form abundant examples of alternating glacial and interglacial sediments. Both aforementioned main Permian warming episodes are very similar to recession phases of glacial sheets.

Causes of climatic changes. Many researchers argue that the northward drift of Pangea with the pole displacement from Antarctica to adjacent ocean was a main (Crowell, 1999; and others) or one of the most

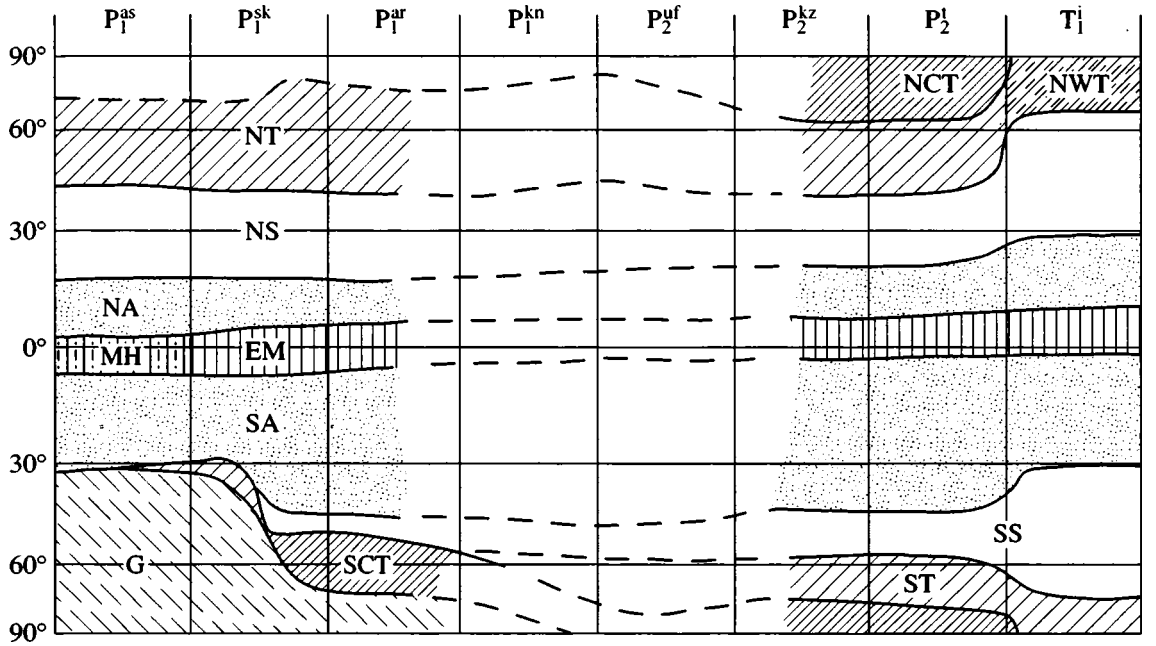

Fig. 3. Evolution of climatic zoning in Pangea through the Permian–Early Triassic. Climatic belts: (G) glacial; (NCT, SCT) northern and southern temperate-cold; (NT, ST) northern and southern temperate; (NWT) northern temperate-warm; (NS, SS) northern and southern semiarid; (NA, SA) northern and southern arid; (EM) equatorial-mountainous arid; (MN) equatorial-mountainous humid (indices P_1^{as}, P_1^{sk}, P_1^{ar}, P_1^{kn}, P_2^{uf}, P_2^{kz}, P_2^{t}, and T_1^{i} denote Permian and Early Triassic ages).

important (Parrish, 1995; Ziegler *et al.*, 1997) causes responsible for deglaciation and, in general, for the Permian–Triassic climatic changes. As is noted in our previous article, such an assumption seems doubtful. According to recently published global reconstructions, Pangea insignificantly drifted northward during the Permian (Golonka *et al.*, 1994; Scotese and Langford, 1995; Ziegler *et al.*, 1997, 1998). Par example, the southern margin of East Antarctica moved away from the South Pole by less than 3° of latitude during the Early Permian and 10° during the entire Permian period. To the contrary, Northeast Asia and North America approached the North Pole. Their northern margins moved from 75 and 48° N to 88 and 58° N, respectively, i.e., approximately by 15°. Most favorable conditions for glaciations development of all glaciation centers were located 15–20° away from the pole (Smith, 1997). Antarctica should remain therefore within this zone, and the insignificant northward drift of the continent could not result in deglaciation and global warming.

In addition, the continental drift is a slow and gradual process, and thus it is an unlikely cause of sudden stepwise and repeated warming events. The small-scale drift of Pangea could likely result in a slight regional cooling in the Northern Hemisphere.[4] Opening of the Neo-Tethys and emergence of a new anticyclonic gyre there, which could be responsible for the additional heat transfer from low to high latitudes, probably had also a regional effect. Naturally, the preceding middle Sakmarian warming and relevant recession of glaciers could not represent an impact of the Middle Permian gyre that could be partly responsible for subsequent gradual warming. Moreover, taking into consideration the aforementioned autonomous functioning of climatic systems in the Earth hemispheres, we can expect that formation of the Neo-Tethyan gyre influenced only northern India and Western Australia.[5]

Other paleogeographic changes recorded in the Permian–Early Triassic should have consequences opposite to warming. Regressions and aridity growth could result only in cooling due to the increase of planetary albedo. The land rise and formation of mountainous systems should also cause cooling because of the negative vertical temperature gradient. Just recollect that climate in Tibet resembles that of plain regions located 10 to 25° of latitude northward in winter and summer, respectively. The land rise in general (Ruddeman and Kutzbach, 1991), and rise of Tibet with the Himalayas in particular (Raymo, 1994), is assumed to be one of main factors responsible for the Late Cenozoic glaciation.

Main factors of Permian–Early Triassic warming were probably related to late Hercynian tectonic and volcanic processes, which changed the atmosphere

[4] This can probably explain the long-lasted cooling in the Sverdrup basin that began in the mid-Permian (Beauchamp, 1995).

[5] Let us remind a more thermophilic composition of fauna in the Indian Province of the Gondwanan Kingdom (Runnegar, 1984).

composition and, as a consequence, resulted in a positive increment of the Earth thermal balance. That increment triggered the chain reaction of processes accelerating and strengthening global warming that became irreversible owing to positive feedbacks in biosphere (Chumakov, 2001). The most important processes were the decrease of the Earth albedo, global degassing of hydrosphere, and decomposition of accumulated gas hydrates. Simultaneously, tectonic and volcanic processes substantially influenced the intensity and composition of the continental runoff and caused the recorded C, O, S, and Sr isotope anomalies.

As was assumed earlier (Stanley, 1989; and others), climatic changes probably played a substantial role in preparation of Permian biotic events, in particularly, of those recorded across the Permian–Triassic boundary (Zharkov and Chumakov, 2001). It seems that the long-lasted aridity increase in Pangea weakened the terrestrial biota, whereas the sudden widening of arid and semiarid belts at the Permian–Triassic boundary time and associated strong warming could bring this biota to the crisis state. The arid climate in and widening of closed drainage areas on land reduced the transport of organic and mineral nutrients to seas. The growing deficiency in these substances enfeebled the marine biota. Stepwise climatic changes, short-term regressions (Valentine, 1973; Ross and Ross, 1995), and sharp reduction of nutrients in seas could disturb feeding chains and result in successive extinctions characterizing the terminal Permian (Leven, 1993; Stanley and Yang, 1994; Erwin, 1995). As is known, the last extinction was of a mass character.

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Eocene Diatoms and Silicoflagellates from the Kronotskii Bay Deposits (East Kamchatka)

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Abstract—Eocene diatoms and silicoflagellates from deposits of the Kronotskii Bay are the oldest known fossils of siliceous phytoplankton from the northwestern Pacific. More than 130 species of 59 diatom genera and 24 species of 5 silicoflagellate genera are identified. Three diatom assemblages of middle Eocene age (those of *Lisitzinia kanayai*, *Lisitzinia inconspicua* var. *triloba*, and *Praecymatosira monomembranaceae* zones) and one presumably corresponding in age to the middle–late Eocene (diatoms of *Rylandsia conniventa* Zone) are established. A diverse silicoflagellate assemblage described for the first time characterizes the *Dictyocha hexacantha* Zone. All the assemblages are supposed to have been formed mainly in bathyal environments and under a relatively high (almost subtropical) temperature of surface water.

Key words: diatoms, silicoflagellates, middle Eocene, Kronotskii Bay, East Kamchatka, paleoenvironment, northwestern Pacific.

The early Paleogene siliceous microplankton is found sporadically in the northern Pacific, because the intense biogenic siliceous sedimentation probably began in the Oligocene time (*Proceedings...*, 1995) to be increased in the Neogene (Baldauf and Barron, 1990). In addition, siliceous skeletons of the Early Paleogene microorganisms are poorly preserved in the marginal basins with bottom sediments of considerable thickness, where heat flow is high, favoring transformation of opal-A into opal-CT (Hein *et al.*, 1978). Finds of Paleogene siliceous microplankton are therefore of a great interest. The region of investigations is also of interest. In publications on the Lower Cenozoic biostratigraphy of East Kamchatka, there are controversial data on age and formation history of deposits in the region (Beniamovskii and Gladenkov, 1996; Levashova *et al.*, 2000). The reconstruction of Cenozoic history is also hampered by absence of reliable data on composition and age of deposits in the East Kamchatka continental slope (Seliverstov, 1998).

This article presents the results of study of diatoms and silicoflagellates from Cenozoic volcanogenic–sedimentary deposits of the Kronotskii Bay. First identifications by E.G. Lupikina, L.M. Dolmatova, and I.B. Tsoy suggested the late Eocene-Oligocene age for the diatom flora from these sediments (Seliverstov, 1998). Glezer *et al.* (1986) and Pushkar' (1987) who described comprehensively the Paleogene diatom assemblages were first to distinguish among them the middle Eocene and late Eocene–Oligocene assemblages. However, different age interpretations are known for assemblages from the same samples. Recent data on stratigraphic ranges of diatoms and newly established species of these fossils revived investigation of Paleogene diatoms from the Kronotskii Bay localities. Based on complex micropaleontological data (diatoms, radiolarians, palynoflora), two constituent parts the Kronotskii sedimentary sequence were distinguished (Tsoy *et al.*, 2000): the upper one with the late Miocene–Pleistocene assemblages and the lower subdivision with predominantly middle Eocene assemblages.

This work is dedicated to description of the Eocene of diatom and silicoflagellate assemblages, their taxonomic composition, age substantiation, and paleoenvironment interpretation included.

BRIEF OUTLINE OF GEOLOGY AND GEOPHYSICAL DATA

The Kronotskii Bay is bordered in the north by the Kronotskii Ridge, a submarine continuation of the synonymous peninsula. The southern end of the ridge is adjacent to the Kamchatka segment of the Kuril–Kamchatka trench. The shelf terrace 10–50 km wide extends into the continental slope to the depth of about 900 m (Seliverstov, 1998). The deep V-shaped structures (the Kronotskii, Ol'ga, and Zhupanovskii canyons) are incised into the acoustic basement the Kronotskii Bay shelf and slope. The largest canyons of the continental slope are as deep as 2 km. An upper part of the sequence is represented by the upper Miocene–Pleistocene volcanogenic-sedimentary deposits showing rhythmical subhorizontal bedding (Tsoy *et al.*, 2000). The maximal thickness (1.5–2 km) of these deposits is measured

in the southern Kronotskii Bay. The lower, acoustically transparent part of sedimentary sequence is relatively persistent in thickness (400–700 m). The bedded and acoustically transparent units are separated by an unconformity that is most distinct in peripheral parts of the depression. The acoustically transparent sediments rest conformably on the acoustic basement composed of volcaniclastic rocks (tuffs, hyaloclastites).

MATERIALS AND METHODS

The sequence was sampled using gravity corer or dredging during Legs 9 and 12 of R/V "Vulkanolog" of the Far East Institute of Volcanology, Russian Academy of Sciences (Table 1; Fig. 1). N.I. Seliverstov headed the expeditions. The dredging intervals were selected based on previously obtained seismoacoustic profiles (Seliverstov, 1998).

The standard procedure and heavy K-Cd solution were used to extract diatom and silicoflagellate remains. Microfossils are identified under magnification of 1350, and valve calculation is performed under magnification of 900. Percentages of species are calculated, depending on abundance rate of fossils, for samples of 100, 200, or 300 specimens. Environmental reconstruction, mostly estimation of paleodepth, is based on proportions between oceanic and neritic species (Jousé, 1962; Koizumi, 1983; Yanagisawa, 1996), and relative temperatures of surface water are estimated based on occurrence frequency of warm-water (low-latitudinal) species of diatoms and silicoflagellates. The Paleogene diatom assemblages are macerated from 9 of 52 samples studied; 12 samples yielded the late Miocene–Pleistocene assemblages, and 31 samples are either barren or contain single unidentifiable remains. The Paleogene assemblages of diatoms and silicoflagellates from the Kronotskii Bay deposits are described below.

Table 1. Sampling sites in the Kronotskii Bay

Site	Latitude, N	Longitude, E	Depth, m
Leg 9 of R/V "Vulkanolog" in 1979			
V9-G4	54°13.8′	161°12.8′	1460
Leg 12 of R/V "Vulkanolog" in 1981			
The Zhupanovskii Canyon			
V12-22	53°34.3′	160°12.2′	887–652
V12-23	53°33.0′	160°11.0′	800–145
V12-24	53°35.0′	160°12.7′	870–720
V12-25	53°30.5′	160°20.4′	2080–1470
V12-26	53°30.7′	160°21.1′	1400–840
The Ol'ga Canyon			
V12-33	54°18.65′	161°08.8′	600–575
V12-34	54°18.3′	161°09.3′	450–215
V12-35	54°19.3′	161°08.3′	577–243
V12-36	54°15.9′	161°10.2′	1120–740
V12-37	54°11.6′	161°11.2′	1210–1186
V12-38	54°11.9′	161°11.4′	1756–1665
The Kronotskii Canyon			
V12-39	53°50.8′	160°43.7′	2703–1817

RESULTS

The Ol'ga Canyon. Sediments of the canyon walls were sampled by means of the gravity corer and dredging (Figs. 1, 2). The upper part of the canyon wall (depth interval 600–215 m, Sites V12-33-V12-35) is mainly composed of small-pebbled conglomerates, tuffaceous diatomites, and siltstones with admixture of pebble and tuffaceous sand (*Nauchno-tekhnicheskii...*, 1980). Diatoms of the Quaternary age were found in the tuffaceous diatomites only. Rocks sampled downward the slope (1120–740 m, Site V12-36) are the weakly lithified siltstone with single pebbles and interbeds of fine-grained sandstones, tuffaceous diatomite (a bed of alternating light-colored tuffaceous diatomite and dark gray mudstone), tuffaceous diatomite with sandstone interbeds and rare pebbles, black volcanomictic poorly sorted sandstone, and tuffaceous gravelstones (*Nauchno-tekhnicheskii...*, 1982). At the Site V12-37 (depth interval 1210–1186 m), black massive siltstones, tuffaceous diatomites, and basalts were recovered. From the depth level of 1460 m (Site V90-D4), the 15-cm-long core of dense tuffaceous diatomite was obtained. The core consists of two different-colored beds, and the lower gray bed (15–9 cm) is in distinct uneven contact with the upper yellow bed (9–0 cm).

In the lower part of the canyon wall (depth interval 1756–1665 m, Site V12-38), the sequence is composed of tuffaceous diatomites alternating with black tuffaceous sandstones and pure sandstones, volcanomictic tuffaceous sandstone, mudstone, volcanogenic pebble, and rounded clasts of tuffaceous diatomites with fucoids. Samples of mudstone (V12-38-3) and tuffaceous sandstone (V12-38-2) contain unidentifiable diatom remains. The tuffaceous diatomites and siltstones yielded diverse siliceous microfossils: diatoms, silicoflagellates, and radiolarians. Four assemblages of diatoms and silicoflagellates have been distinguished.

The diatom assemblage (25 species) from Sample V12-36-1-4 of dense tuffaceous diatomite consists of abundant *Lisitzinia kanayai* (Fenner) Gleser and associated *Riedelia borealis* Sheshukova, *Paralia crenulata* (Grunow) Gleser, *Hemiaulus polycystinorum* Ehrenberg, *H. polymorphus* Grunow, *Stephanopyxis* spp., *Lisitzinia inconspicua* var. *trilobata* Gleser, *Azpeitia tuberculata* var. *atlantica* (Gleser et Jousé) Sims, *Coscinodiscus decrescens* Grunow, and other species (Table 2). Percentages of different ecological groups (55.5% of oceanic, 24% of neritic, and 1% of benthic species) indicate a bathyal habitat of this flora. Sili-

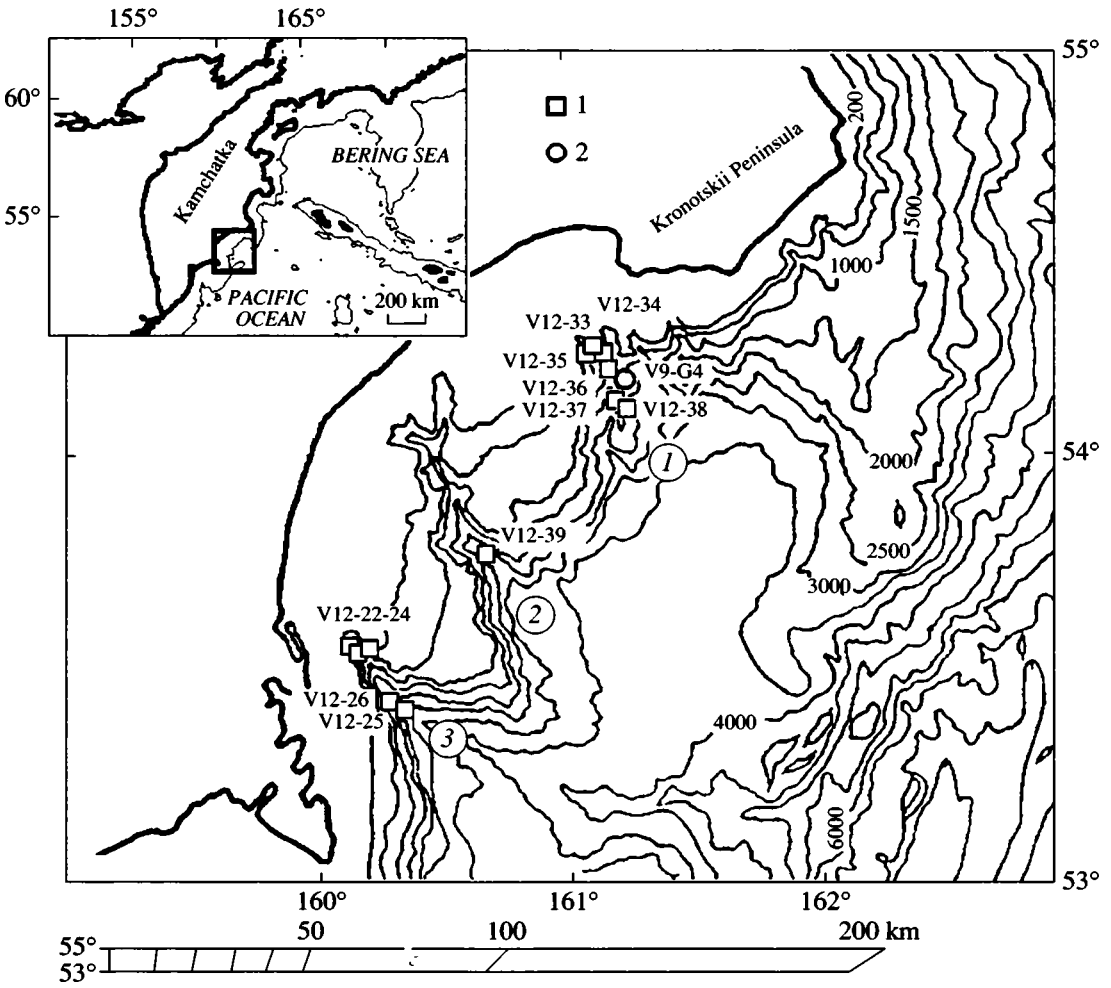

Fig. 1. Sampling sites in the Kronotskii Bay: (1) dredging site, (2) gravity core site; encircled numbers denote the Ol'ga (*1*), Kronotskii (*2*), and Zhupanovskii (*3*) canyons (bathymetry after Gnibidenko *et al.*, 1983).

coflagellates are represented by the middle Eocene warm-water *Naviculopsis foliacee* Deflandre.

Similar diatom assemblages from samples of siltstone (V12-36-1-1) and tuffaceous diatomite (V12-38-1) are dominated by different species (*Lisitzinia inconspicua* var. *trilobata* Gleser versus *Paralia crenulata* (Grunow) Gleser respectively). Both assemblages include diverse *Hemiaulus* and *Stephanopyxis* forms associated with *Azpeitia tuberculata* var. *atlantica* (Gleser et Jousé) Sims, *Asterolampra vulgaris* Greville, *Coscinodiscus decrescens* Grunow, *Navicula udintsevii* Schrader, *Praecymatosira monomembranaceae* (Schrader) Strelnikova, *Coscinodiscus tenerrimus* Jousé, *C. hajosiae* Fenner, *Costopyxis trochlea* (Hanna) Strelnikova, *Pterotheca aculeifera* Grunow, *Peponia* sp. and others (Table 2). The assemblages are of different ecological affinity: that from Sample V12-36-1-1 shows high percentages of the oceanic (45.3%) and warm-water (40.7%) species, whereas the other one from Sample V12-38-1 includes about 56.9% of neritic species. Rare silicoflagellates *Corbisema hastata globulata* Bukry, *Dictyocha deflandrei* Frenguelli ex Gleser, and *Naviculopsis foliacee* Deflandre were found in the latter sample only.

The tuffaceous diatomite beds sampled by gravity coring (Site V9-G4, depth 1460 m) bear two diatom assemblages. The assemblage of the lower bed (samples V9-G4-2947 and V9-G4-2949) consists of 48 species. Its obvious dominant is *Paralia crenulata* (Grunow) Gleser occurring as well preserved individual valves and as fragments of colonies. Stratigraphically important species of the assemblage are *Peponia barbadense* Greville, *Peponia* sp., *Praecymatosira monomembranaceae* (Schrader) Strelnikova, *Coscinodiscus hajosiae* Fenner, *Distephanosira architecturalis* (Bryn) Gleser, *Coscinodiscus* cf. *excavatus* Castracane, *Navicula udintsevii* Schrader, *Azpeitia tuberculata* var. *atlantica* (Gleser et Jousé) Sims, *Lisitzinia inconspicua* var. *inconspicua* Gleser, *L. inconspicua* var. *trilobata* Gleser and others (Table 2). Silicoflagellates are represented by *Corbisema triacantha* (Ehrenberg) Bukry et Foster and *Distephanus* sp. The assemblage demon-

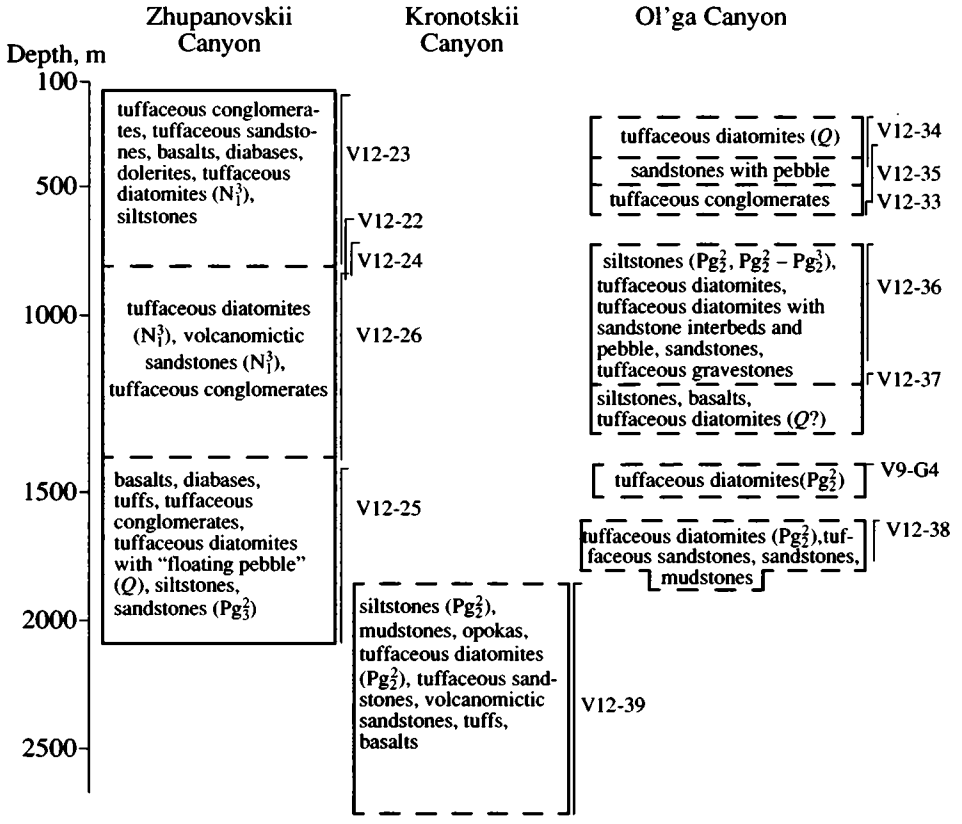

Fig. 2. Lithology and age (based on diatoms and silicoflagellates) of sequence intervals sampled in the Kronotskii Bay canyons, East Kamchatka; (V9-G4, V12-22-V12-39) site numbers; age indices: (Pg_2^2) middle Eocene, ($Pg_2^2 - Pg_2^3$) middle–upper Eocene. (Pg_3^2) upper Oligocene, (N_1^3) upper Miocene, (Q) Pleistocene.

strates the distinct predominance (71–77%) of neritic species over the oceanic (12–20%) and less abundant warm-water (6–10%) forms.

The diatom assemblage from the upper bed (Sample V9-G4-2948) is similar though less diverse (25 species) and lacking dominating species. Its characteristic taxa are abundant *Hemiaulus* forms, *Paralia crenulata* (Grunow) Gleser, *Proboscia interposita* (Hajós) Jordan et Priddle, *Distephanosira architecturalis* (Brun) Gleser, and *Praecymatosira monomembranaceae* (Schrader) Strelnikova. Characteristic of the assemblage is predominance of oceanic species (45.6%), reduced content of neritic forms (39.2%), and increased percentage of warm-water taxa (about 30%).

The layered siltstone sample (V12-36-1-2) is remarkable owing to high abundance and diversity of diatoms and silicoflagellates, species composition of which is peculiar. The diatom assemblage lacking dominants consists of 81 species, mostly of abundant and diverse representatives of *Hemiaulus, Azpeitia, Coscinodiscus, Asterolampra*, and *Chaetoceros* genera. It includes the following species known predominantly from the middle–upper Eocene deposits of different latitudes: *Hemiaulus incisus* Hajós, *Pyxilla gracilis* Tempére et Forti, *Trinacria excavata* Heiberg, *Rylandsia conniventa* Gleser, Dolmatova et Lupikina, *Coscinodiscus gombosii* Gleser, Dolmatova et Lupikina, *Kisseleviella cuspidata* Gleser, Dolmatova et Lupikina, *Coscinodiscus hajosiae* Fenner, *Pinnularia* aff. *antiqua* Tscheremissinova, *Asterolampra vulgaris* Greville, *A. praeacutiloba* Fenner, *A. schmidtii* Hajós, *Pseudotriceratium radiosoreticulatum* Grunow, *Pseudopodosira corolla* (A. Schmidt) Hajós, *Thalassiosira dubiosa* Schrader, *Sceptroneis vermiformis* Schrader, *Distephanosira architecturalis* (Brun) Gleser, and others. Taxa of the assemblage represent oceanic (59.2%), neritic (28.8%), and benthic (6%) forms. There is a single fresh-water form *Aulacoseira* sp. among them. Percentage of supposed warm-water species is high (44%).

The sample yielded abundant and diverse silicoflagellates (24 species) representing marine unicellular algae with siliceous skeletons. These are *Naviculopsis foliaceae* Deflandre, *Corbisema hastata globulata* Bukry, C. apiculata (Lemmermann) Hanna, *Bachmannocena apiculata inflata* Bukry, *B. paulschulzii* Bukry, *Dictyocha hexacantha* Schulz, *D. spinosa* (Deflandre) Gleser, *D. deflandrei* Frenguelli ex Gleser and others (Table 3).

Table 2. Diatoms from the Kronotskii Bay deposits

Diatom assemblage		*L. kanayai*		*L. inconspicua* v. *trilobata*		*P. monomembranaceae*			*R. conniventa*
Taxonomic composition	Ecology	V12-39-1-4	V12-36-1-4	V12-36-1-1	V12-38-1	V9-G4-2947	V9-G4-2949	V9-G4-2948	V12-36-1-2
Actinocyclus ingens Rattray	pow				3.3				*
Actinoptychus senarius Ehrenberg	b		0.5	0.3	1.3	0.3			2.0
Actinoptychus sp.	b				1.0	0.3		0.3	
Anaulus sp.	b								*
Arachnoidiscus indicus Ehrenberg	b	1.0						0.3	0.4
Arachnoidiscus spp.	b	1.0		0.3	0.3	0.3	0.3		*
Asterolampra insignis A. Schmidt	pow					0.3			0.4
Asterolampra marylandica Ehrenberg	pow								*
Asterolampra praeacutiloba Fenner	pow								0.2
Asterolampra punctifera (Grunow) Hanna	pow								0.4
Asterolampra schmidtii Hajós	pow								0.4
Asterolampra spp.	pow								0.4
Asterolampra vulgaris Greville	pow	1.0		0.3					0.4
Aulacodiscus cf. *lahusenii* Witt	b								*
Aulacodiscus inflatus var. *spinifer* Brun	b								0.2
Aulacodiscus lahusenii Witt	b								0.2
Aulacoseira sp.	fw								0.2
Azpeitia (Coscinodiscus) gombosii Gleser, Dolmatova et Lupikina	pow						0.3	0.3	6.0
Azpeitia oligocenica (Jousé) Sims	pow						0.3		0.2
Azpeitia sp.	pow					0.3			
Azpeitia tuberculata var. *atlantica* (Gleser et Jousé) Sims	pow		0.5	8.7	0.6		0.3	0.7	1.6
Biddulphia sp.	b	1.0			0.3				
Biddulphia tuomei (Bailey) Roper	b	1.0		0.3	0.3				
Bipalla (*Melosira*) *oamaruensis* (Grove et Sturt) Gleser	pn			0.3					
Brightwellia sp. (*B.* cf. *imperfecta* Jousé)	pow								0.2
Cavitatus cf. *jouseanus* Sheshukova	pow							0.3	0.2
Cestodiscus spp.	pow				0.3				*
Chaetoceros (*Xanthiopyxis*) *panduraeformis* (Pantocsek) Gombos	s		0.5			0.3			0.2
Chaetoceros spp.	s	2.0	0.5		2.0		0.3	0.7	8.0
Clavicula polymorpha Grunow et Pantocsek	b			0.3					*
Coscinodiscus aff. *excavatus* Castracane	pow					0.3	0.3	0.3	
Coscinodiscus apiculatus var. *ambiguus* Grunow	po				0.3				
Coscinodiscus argus Ehrenberg	po	1.0			0.3		1.3		0.4
Coscinodiscus asteromphalus Ehrenberg	pow								0.4
Coscinodiscus decrescenoides Jousé	po	1.0							
Coscinodiscus decrescens Grunow	po	1.0	4.0	2.7	0.3		0.3		10.0
Coscinodiscus hajosiae Fenner (=*Hyalopoda spiralis* (Hajós) Kozyrenko et Jackovschikova, Strel'nikova *et al.*, 1998)	po				0.3		0.3	0.3	2.0
Coscinodiscus marginatus Ehrenberg	po	1.0			0.7	0.7	3.0	1.7	
Coscinodiscus mirabilis Jousé	po	1.0			0.3				*
Coscinodiscus monicae Grunow	po								*
Coscinodiscus oculus iridis Ehrenberg	po	1.0		0.3			0.3		0.4
Coscinodiscus sectoralis Gleser, Dolmatova et Lupikina	p								0.8

Table 2. (Contd.)

Diatom assemblage		*L. kanayai*		*L. inconspicua* v. *trilobata*		*P. monomembranaceae*			*R. conniventa*
Taxonomic composition	Ecology	V12-39-1-4	V12-36-1-4	V12-36-1-1	V12-38-1	V9-G4-2947	V9-G4-2949	V9-G4-2948	V12-36-1-2
Coscinodiscus sp. A (sensu Barron, Mahood, 1993)	p			0.3					
Coscinodiscus spp.	p	6.0	1.5	0.7	4.7	0.3	5.7	6.0	0.8
Coscinodiscus subtilis Ehrenberg	p						0.3		*
Coscinodiscus tenerrimus Jousé	p			1.3					0.6
Costopyxis schulzii (Steinecke) Gleser	s				0.7				
Costopyxis trochlea (Hanna) Strelnikova	s				1.0				
Craspedodiscus (*Porodiscus*) *splendidus* (Greville) Gombos	pow				0.3				
Craspedodiscus klavsenii Gründler	pow								*
Craspedodiscus moelleri A. Schmidt	pow	1.0							
Cymatosira spp.	p					0.3			0.2
Distephanosira (*Melosira*) *architecturalis* (Brun) Gleser	pn	1.0		0.3		0.3		10.3	0.8
Drepanotheca (*Eunotogramma*) *bivitata* (Grunow et Pantocsek) Schrader	b					0.3			
Endictya spp.	po			0.3			0.3		*
Entopyla frickei Hanna	b	1.0							0.2
Ethmodiscus sp.	po								*
Eupodiscus cf. *oamaruensis* Grunow	p								0.2
Genus et species indet. 1 (sensu Gombos, 1983)	p								*
Genus et species indet. 2 (sensu Gombos, 1983)	p			0.3					
Hemiaulus cf. *vesicarius* Strelnikova	p				1.0				
Hemiaulus incisus Hajós	pow								4.8
Hemiaulus polycystinorum Ehrenberg	pow	1.0	8.0	10.2	1.7	3.3	0.3	24.0	6.0
Hemiaulus polymorphus Grunow	pow	1.0	10.0	10.2	2.0	0.7	0.7	2.0	6.0
Hemiaulus polymorphus v. *frigida* Grunow	pow				0.7				
Hemiaulus spp.	pow	5.0	0.5	9.3	5.3	2.3	2.7	1.3	8.0
Hemiaulus subacutus Fenner	pow				0.7				
Hyalodiscus scoticus (Kutzing) Grunow	pn				0.3				
Hyalodiscus spp.	pn	2.0		0.3	0.3		0.3		0.2
Istmia sp.	b	1.0							
Kisseleviella cuspidata Gleser, Dolmatova et Lupikina	pn								0.8
Liradiscus ovalis Greville	s	1.0							6.2
Lisitzinia brachiatum (Brightwell) Gleser	p				*				
Lisitzinia inconspicua (Greville) Gleser var. *inconspicua* Gleser	p			0.3	0.3	1.0	0.3	0.7	
Lisitzinia inconspicua (Greville) Gleser var. *trilobata* Gleser	p	*	1.0	30.6	11.3	0.3			
Lisitzinia kanayai (Fenner) Gleser	pw	34.0	24.5	0.3					
Lisitzinia sp. (*Triceratium* sp. sensu Barron *et al.*, 1984)	p		0.5						
Navicula udintsevii Schrader	p			0.3	0.3	0.3	0.3	0.3	
Odontella fimbriata (Greville) Schrader	p			0.3					
Odontotropis carinata Grunow ?	pn			0.3					
Odontotropis sp.	pn								1.2
Paralia crenulata (Grunow) Gleser	pn		7.0	0.3	41.3	73.3	66.7	22.3	
Paralia polaris (Grunow) Gleser	pn	2.0		0.3	0.7		0.3		*
Paralia sulcata (Ehrenberg) Cleve	pn	1.0			0.3			0.3	1.6
Peponia barbadense Greville	pn					*			
Peponia sp. (=*Peponia* sp. 1 sensu Fenner, 1978)	pn				0.3	0.7	0.3	1.7	
Pinnularia aff. antiqua Tscheremissinova	b								0.2

Table 2. (Contd.)

Diatom assemblage		*L. kanayai*		*L. inconspicua* v. *trilobata*		*P. monomembranaceae*			*R. conniventa*
Taxonomic composition	Ecology	V12-39-1-4	V12-36-1-4	V12-36-1-1	V12-38-1	V9-G4-2947	V9-G4-2949	V9-G4-2948	V12-36-1-2
Praecymatosira monomembranaceae (Schrader) Strelnikova	p			5.0		7.3	2.0	7.0	
Proboscia cf. *interposita* (Hajós) Jordan et Priddle	po	1.0				1.7	5.7	8.0	2.0
Proboscia interposita (Hajós) Jordan et Priddle	po	1.0	2.0	1.3	0.7	1.3	3.0	6.7	*
Pseudopodosira corolla (A. Schmidt) Hajós	pn								*
Pseudopyxilla americana (Ehrenberg) Forti	s					0.3			
Pseudopyxilla sp.	s				0.3				
Pseudotriceratium radiosoreticulatum Grunow	pn			0.3					1.6
Pterotheca aculeifera Grunow	s				1.7				
Pterotheca danica Grunow	s	1.0	0.5			0.3	0.3		
Pyxilla gracilis Tempère et Forti	pn				0.3				1.6
Pyxilla spp.	pn	1.0		0.3	0.3	0.3			0.4
Rhizosolenia spp.	po								0.4
Riedelia borealis Sheshukova	p	5.0	14.0	0.3	1.7				
Riedelia claviger (A. Schmidt) Schrader et Fenner	p								*
Riedelia pacifica Jousé	p		1.0	0.3					
Riedelia sp. 1 (sensu Schrader, Fenner, 1976)	p		5.0	0.3					*
Rutilaria spp.	b							0.3	0.2
Rylandsia biradiata Greville	pow	1.0				*	0.3		
Rylandsia conniventa Gleser, Dolmatova et Lupikina	pow								2.4
Sceptroneis pesplanus Schrader	b			0.3		0.3			
Sceptroneis spp.	b								0.4
Sceptroneis tenue Schrader et Fenner	b				0.3				*
Sceptroneis vermiformis Schrader	b								1.0
Stellarima microtrias (Ehrenberg) Hasle et Sims	pow	3.0	1.0	2.0	0.3				7.2
Stellarima stellaris (Roper) Hasle et Sims	pow	1.0				0.3	0.3		*
Stephanogonia sp.	s				0.3				
Stephanopyxis cf. *aciculatus* Dolmatova	pn								*
Stephanopyxis cf. *broschii* Grunow	pn								0.4
Stephanopyxis cf. *superba* Grunow	pn	1.0							
Stephanopyxis ferox (Greville) Ralfs	pn		2.0						
Stephanopyxis grunowii Grove et Sturt	pn	1.0	2.0	0.3	0.3	0.3			
Stephanopyxis marginata Grunow	pn	2.0	2.0	0.7	0.7	0.3	0.3	0.3	*
Stephanopyxis spp.	pn	10.0	10.0	4.3	5.7	0.7	2.0	3.7	0.4
Stephanopyxis turris (Greville et Arnott) Ralfs	pn	1.0	1.0	4.3	1.0	0.3	0.3		5.6
Stephanopyxis turris var. *intermedia* Grunow	pn				0.3				
Stictodiscus hardmanianus Greville	b					0.3			0.2
Stictodiscus spp.	b	1.0	0.5		1.0				
Thalassiosira dubiosa Schrader	p								0.8
Thalassiosiropsis wittiana (Pantocsek) Hasle	p								*
Triceratium arcticum Brightwell	b						0.3		
Trinacria excavata Heiberg	p								1.6
Total, %		100	100	100	100	100	100	100	100

Note: (p) planktonic, (b) benthic, (n) neritic, (o) oceanic, (w) warm-water diatoms, and (s) spores; taxa identified after counts are indicated by asterisk; ecological affinity is in accord data of Sheshukova-Poretskaya (1967), Baldauf and Barron (1987), and Fenner (1985).

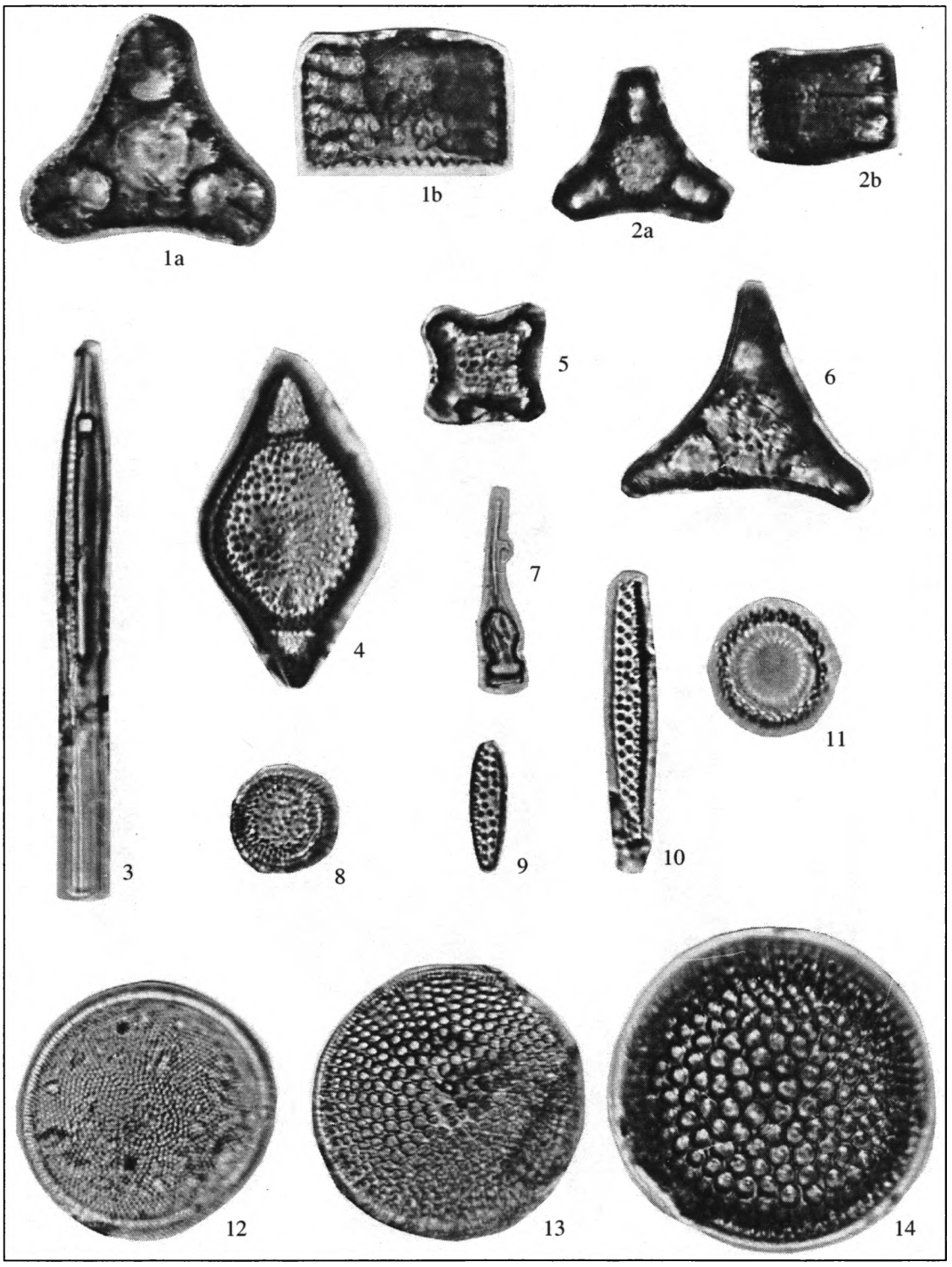

Plate 1. Characteristic species of the middle Eocene diatom assemblages from the Kronotskii Bay deposits (East Kamchatka)

(1) *Lisitzinia kanayai* (Fenner) Gleser: (1a) valve view, (1b) belt view, ×1300; (2): *Lisitzinia inconspicua* (Greville) Gleser, var. *triloba* Gleser: (2a) valve view, (2b) belt view, ×1300; (3) *Riedelia borealis* Sheshukova, ×1300; (4) *Peponia* sp., ×1300; (5) *Lisitzinia inconspicua* var. *inconspicua* Gleser, ×1300; (6) *Lisitzinia brachiatum* (Brightwell) Gleser, ×1300; (7) *Pterotheca aculeifera* Grunow, ×650; (8) *Distephanosira architecturalis* (Brun) Gleser, ×1300; (9, 10) *Praecymatosira monomembranaceae* (Schrader) Strelnikova, ×1300; (11) *Paralia crenulata* (Grunow) Gleser, ×1300; (12) *Coscinodiscus hajosiae* Fenner; (13) *Azpeitia tuberculata* var. *atlantica* (Gleser et Jousé) Sims; (14) *Coscinodiscus decrescens* Grunow.

Specimens: (1a, 1b, 3) from Sample V12-39-1-4; (2a, 2b, 6, 7, 12) from Sample V12-38-1; (4, 5, 8, 10, 11, 14) from Sample V9-G4-2949; (9) from Sample V9-G4-2948; (13) from Sample V12-36-1-1.

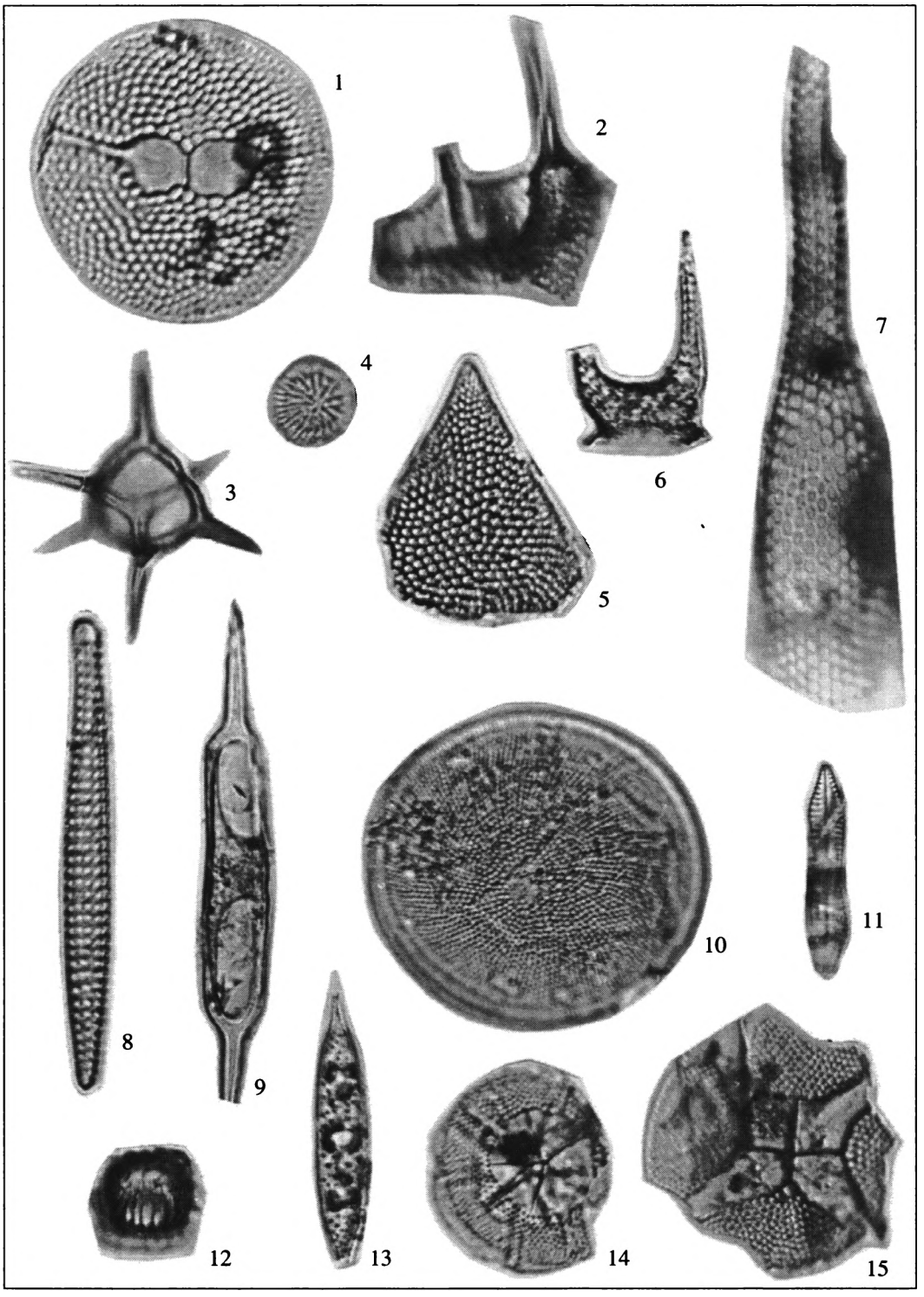

Plate 2. Characteristic diatom species of the middle (?)-late Eocene and silicoflagellates from the Kronotskii Bay deposits
(1) *Rylandsia conniventa* Gleser, Dolmatova et Lupikina, ×1300; (2) Genus et species indet. 1 (sensu Gombos, 1983), ×1500; (3) *Dictyocha hexacantha* Schulz, ×1500; (4) *Thalassiosira dubiosa* Schrader, ×1500; (5) *Pseudotriceratium radiosoreticulatum* Grunow, ×1300; (6) *Hemiaulus incisus* Hajós, ×1300; (7) *Pyxilla gracilis* Tempére et Forti, ×1300; (8) *Sceptroneis vermiformis* Schrader, ×1300; (9) *Naviculopsis foliaceae* Deflandre, ×600; (10) *Coscinodiscus hajosiae* Fenner, ×1300; (11) *Pinnularia* aff. *antiqua* Tscheremissinova, ×1500; (12) *Costopyxis trochlea* (Hanna) Strelnikova, ×1300; (13) *Kisseleviella cuspidata* Gleser, Dolmatova et Lupikina, ×1300; (14) *Asterolampra punctifera* (Grunow) Hanna, ×1300; (15) *Asterolampra praeacutiloba* Fenner, ×1000.
All specimens are from Sample V12-36-1-2.

The silicoflagellate assemblage is dominated by the warm-water *Naviculopsis foliaceae* Deflandre and includes diverse and abundant *Corbisema* species, which are also suspected to be of the warm-water origin (Bukry, 1987) because of their abundance in the low latitudes. The high content of *Dictyocha hexacantha* Schulz is also characteristic of the low and, to a lesser extent, of moderate latitudes.

The Kronotskii Canyon. The lower part of the sequence (Site V12-39, depth interval 2703–1817 m) is composed here of volcanomictic and tuffaceous sandstones, coarse-grained tuffs, basalts, tuffaceous diatomites, carbonaceous siltstones, opokas, and siliceous mudstones. Diatoms were found in the siltstones (sample V12-39-1-4). Other rocks are either barren or contain single unidentifiable remains.

The diatom assemblage is analogous of that of the Ol'ga Canyon (Sample V12-36-1-4), dominated by zonal species *Lisitzinia kanayai* (Fenner) Gleser of the middle Eocene. It also includes less frequent though diverse of *Hemiaulus* and *Stephanopyxis* forms associated with single *Lisitzinia inconspicua* var. *trilobata* Gleser, *Asterolampra vulgaris* Greville, *Coscinodiscus decrescens* Grunow, *Distephanosira architecturalis* (Brun) Gleser, *Riedelia borealis* Sheshukova and others (Table 2). As in the Ol'ga Canyon, the assemblage predominantly consists of oceanic (57%) and low-latitude (47%) species, which suggest similar paleoenvironments for both assemblages. Silicoflagellates are represented by *Dictyocha deflandrei* Frenguelli ex Gleser. According to V.V. Shastina, associated radiolarians constitute the *Artobotrus auriculaleporis* assemblage (Tsoy *et al.*, 2000). This assemblage is correlative with radiolarians of the middle Eocene *Artobotrus norvegiensis* Zone and of the middle–upper Eocene *Phacodiscus testatus grandis* Beds of the Norwegian Sea.

The Zhupanovskii Canyon. The sampled interval was the upper part of the section (Sites V12-22, V12-23, V12-24, and V12-26, depth interval 1400–145 m) mainly represented by tuffaceous diatomites, tuffaceous conglomerates, volcanomictic sandstones, and siltstones. The tuffaceous diatomites and sandstones yielded diatom assemblages of the late Miocene–Pliocene, some of which contain the reworked Paleogene species. The deeper interval (2080–1470 m) was dredged at the Site V12-25. The recovered samples represent basalts, tuffs, tuffaceous conglomerates, tuffaceous diatomites, siltstones, and sandstones. The tuffaceous diatomites contain diatoms of Quaternary age. Sandstone Sample 25-3-2 yielded the upper Eocene diatom assemblage of the *Stephanopyxis marginata-Goniothecium decoratum* Zone (Pushkar', 1987). However, the assemblage may be reworked because typical middle Eocene species (e.g., *Triceratium barbadense* Greville, *T. mirabile* Jousé, *Sheshukovia (Triceratium) inconspicua* Gleser, *Praecymatosira monomembranaceae* (Schrader) Strelnikova) are associated with characteristic Oligocene forms (e.g., *Lisitzinia ornata* Jousé, *Coscinodiscus vigilans* Schmidt, *Actinocyclus* sp.). Popova (1989) described the middle Eocene radiolarian assemblage from the Zhupanovskii Canyon, but she did not indicate the site and sample numbers, and stratigraphic allocation of the assemblage thus remains ambiguous.

DISCUSSION

Investigated deposits of the Kronotskii Bay yielded diatom assemblages of the Paleogene age. They characterize lower parts of the largest submarine Ol'ga and Kronotskii canyons.

Assemblage 1 is macerated from tuffaceous diatomites and siltstones (Samples V12-39-1-4, V12-36-1-4). Its dominating species is *Lisitzinia kanayai* (Fenner) Gleser. The assemblage contains *Riedelia borealis* Sheshukova, *Coscinodiscus decrescens* Grunow, *Hemiaulus polycystinorum* Ehrenberg, *H. polymorphus* Grunow, *Lisitzinia inconspicua* var. *trilobata* Gleser, *Rylandsia biradiata* Greville, *Asterolampra vulgaris* Greville, *Azpeitia tuberculata* var. *atlantica* (Gleser et Jousé) Sims, *Craspedodiscus moelleri* A. Schmidt, *Distephanosira architecturalis* (Brun) Gleser, *Entopyla frikei* Hanna, and others.

According to presence of abundant *Lisitzinia (Triceratium) kanayai*, the assemblage is correlated with diatoms of synonymous zone established in low latitudes (Fig. 3). According to Fenner (1984), this zone corresponds to calcareous nannoplankton zones NP14-NP15 of the lowermost middle Eocene (lower Lutetian). The assemblage includes *Rylandsia biradiata* Greville and *Lisitzinia inconspicua* var. *trilobata* Gleser, which have their first occurrences in this zone. The zonal species domination makes the assemblage similar to that from the Kellogg Shale of northern California, where this subdivision is dated back to the middle Eocene and corresponds to zone P12 of planktonic foraminifers and to the interval CP13c-CP14a of nannoplankton zonation (42.0–45.0 Ma, upper Lutetian–lower Bartonian; Barron *et al.*, 1984). Radiolarian faunas of the Kronotskii Canyon and Kellogg Shale also include some species in common (Tsoy *et al.*, 2000). The low-latitude oceanic species dominating in diatom assemblage are indicative of relatively deep-water (bathyal?) and warm (nearly subtropical) environments.

Assemblage 2 is established in tuffaceous diatomites and siltstones (Samples V12-38-1, V12-36-1-1). Being similar to the previous one, this assemblage is dominated by another typical Eocene species *Lisitzinia inconspicua* var. *triloba* Gleser and includes *Lisitzinia inconspicua* var. *inconspicua* Gleser, *Navicula udintsevii* Schrader, *Costopyxis trochlea* (Hanna) Strelnikova, *Pyxilla gracilis* Tempére et Forti, *Riedelia pacifica* Jousé, *Coscinodiscus tenerrimus* Jousé, and others. Similar diatom assemblages are characteristic of volcanogenic–sedimentary Ushchel'e sequence, upper

Table 3. Silicoflagellates from the Kronotskii Bay deposits

Silicoflagellates	V12-39-1-4	V12-36-1-4	V12-36-1-1	V12-38-1	V9-G4-2947	V9-G4-2949	V9-G4-2948	V12-36-1-2
Bachmannocena apiculata inflata Bukry								16
Bachmannocena paulschulzii Bukry								7
Corbisema apiculata (Lemmermann) Hanna								1
Corbisema glezerae Bukry								8
Corbisema hastata globulata Bukry				1				20
Corbisema hastata hastata (Lemmermann) Bukry								1
Corbisema lamillifera (Gleser) Bukry								1
Corbisema ovalis Perch-Nielsen								1
Corbisema sp.								1
Corbisema triacantha (Ehrenberg) Bukry et Foster						1		1
Dictyocha spinosa (Deflandre) Gleser								1
Dictyocha deflandrei Frenguelli ex Gleser	1			1				1
Dictyocha frenguellii Deflandre								1
Dictyocha hexacantha Schulz								10
Dictyocha pentagona (Schulz) Bukry et Foster								4
Dictyocha sp. (asperoid) sensu Bukry, 1987								1
Dictyocha sp.								10
Distephanus cf. *bolivinensis bolivinensis* (Frenguelli) Bukry								2
Distephanus crux (Ehrenberg) Haeckel				1				1
Distephanus quinquangellus Bukry et Foster								4
Distephanus sp.						1		
Naviculopsis biapiculata (Lemmerman) Frenguelli								3
Naviculopsis constricta (Schulz) Frenguelli								1
Naviculopsis foliacee Deflandre		1		2				55
Total	1	1		5		2		151

Kubovaya Subformation, and Kozlovskaya Formation of the Kronotskii Peninsula (*Strukturno-veshchestvennye...*, 1995). Based on planktonic and benthic foraminifers and nannoplankton, these subdivisions are referred mostly to the Bartonian *Reticulofenestra umbilica* and NP16-NP17 zones of the middle Eocene (*Strukturno-veshchestvennye...*, 1995; Shcherbinina, 1997). Diatom assemblages from the on-land formations and from the Ol'ga of the Kronotskii Bay canyons include the following species in common: *Coscinodiscus argus* Ehrenberg, *C. monicae* Rattray, *Sheshukovia* (=*Lisitzinia*) *inconspicua* var. *triloba* Gleser, *Coscinodiscus decrescens* Grunow, *Hemiaulus polymorphus* Grunow, *H. polycystinorum* Ehrenberg, *Riedelia borealis* Sheshukova, *Actinocyclus ingens* Rattray, and others. V.V. Shastina also noted some common elements in corresponding radiolarian faunas (Tsoy *et al.*, 2000).

The assemblage under discussion includes some species typical of the middle Eocene *Trinacria excavata* f. *tetragona* and *Craspedodiscus oblongus* zones of the Norwegian Sea (Dzinoridze *et al.*, 1978; Fenner, 1985). It also shows similarity to the assemblage of non-formal *Lisitzinia (Triceratium) inconspicua* var. *trilobata* Gleser Zone of the uppermost middle Eocene, which was established in the Bateque Formation of California (McLean and Barron, 1988). The zonal species occurs in abundance in the upper part of the middle Eocene Kellogg section of California. Some species of the assemblage are present in the middle Eocene assemblage of the Kreyenhagen Shale (Oro Loma section) of the western San Joaquin valley, California (McLean and Barron, 1988) and in the South Atlantic (Fenner, 1978). The Bateque Formation is referred to the uppermost middle Eocene (upper subzone CP14a to lower subzone CP14b). These data suggest the terminal middle Eocene age of Assemblage 2. The inferred younger age of the assemblage is consistent with data on associated radiolarians, which are characteristic of

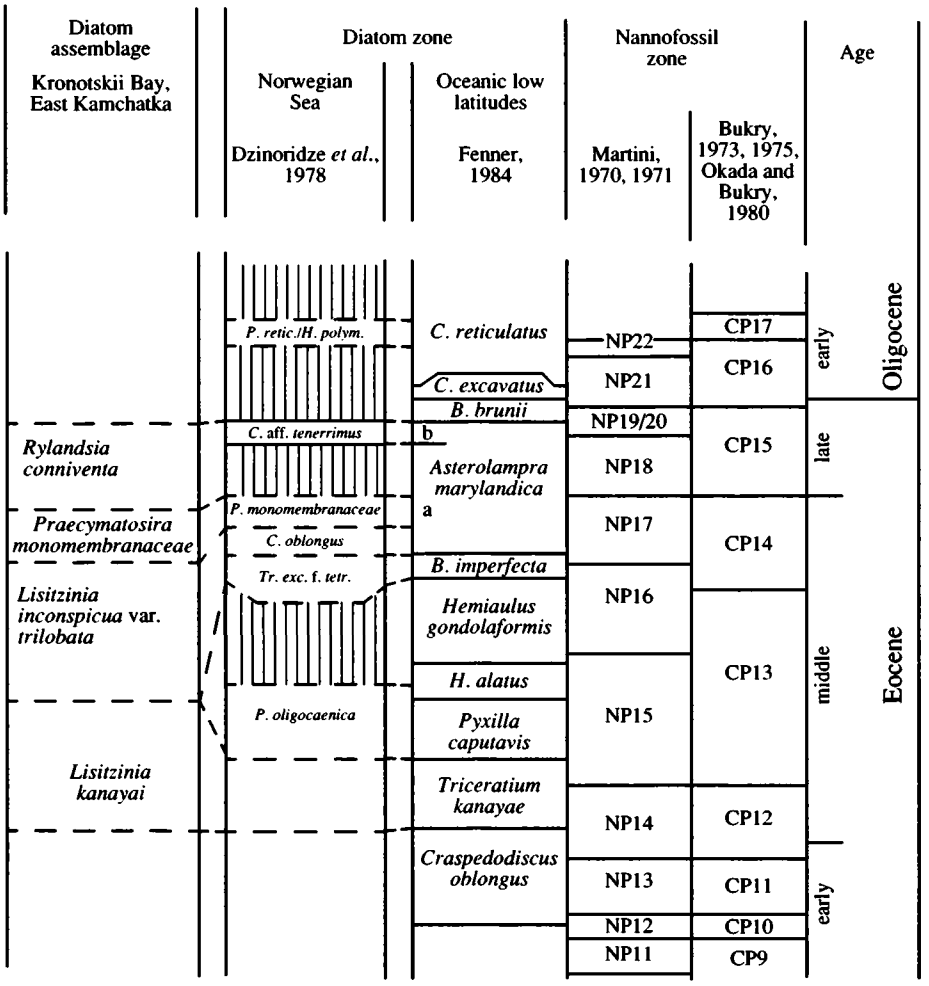

Fig. 3. Of the Diatom assemblages of the Kronotskii Bay (East Kamchatka) correlated with diatom zones of the Norwegian Sea and oceanic low latitudes, and with nannoplankton zonation.

the upper Eocene and even Oligocene deposits (Tsoy *et al.*, 2000). Assemblage 2 was formed in cooler environments than Assemblage 1, as it is evident from abundance of *Lisitzinia inconspicua* var. *triloba*, which is widespread in high latitudes of the Norwegian Sea (Barron *et al.*, 1984). Assemblage 2 contains a greater amount (up to 57%) of neritic species.

Assemblage 3 (samples V90-G4-2947, V9-G4-2948, V9-G4-2949) includes the zonal species *Praecymatosira monomembranaceae* (Schrader) Strelnikova associated with *Peponia barbadense* Greville, *Peponia* sp., *Coscinodiscus hajosiae* Fenner, *Distephanosira architecturalis* (Brun) Gleser, *Navicula udintsevii* Schrader, *Rylandsia biradiata* Greville, and others. The dominant is *Paralia crenulata* (Grunow) Gleser occurring in abundance in the Paleogene deposits of the Urals and southern Kazakhstan (Jousé, 1978). Closely related forms characterize the upper Eocene beds of the equatorial Atlantic (Glezer and Jousé, 1974). The zonal species *Praecymatosira monomembranaceae* (Schrader) Strelnikova was described from the middle Eocene of the Vöring Plateau of the Norwegian Sea (Schrader and Fenner, 1976; Dzinoridze *et al.*, 1978; *Istoriya...*, 1979; Goll, 1989) and from the southern Atlantic (Hajós, 1976). The assemblage with *Praecymatosira monomembranaceae* (= *Cymatosira* sp. B) characterizes the synonymous zone of the Norwegian Sea (Dzinoridze *et al.*, 1978). According to Fenner (1985), this zone corresponds in scope to the *Trinacria excavata* f. *tetragona* and *Coscinodiscus oblongus* zones of the lowermost middle Eocene. Another characteristic species *Peponia* sp. was previously found only in the middle Eocene beds of the San Paulo Plateau of the South Atlantic (Fenner, 1978). Strel'nikova (1992) attributed the zone to the lower part of the middle Eocene.

According to presence of zonal species *Praecymatosira monomembranaceae* (Schrader) Strelnikova and some taxa characteristic of the synonymous zone (e.g., *Lisitzinia inconspicua* Gleser, *Coscinodiscus hajosiae* Fenner) and of the middle Eocene strata in high and low latitudes, Assemblage 3 can be attributed to the *Praecy-*

matosira monomembranaceae Zone of the uppermost middle Eocene. The assemblage composition reflects a transition from neritic to bathyal environments.

Assemblage 4 from siltstone Sample C12-36-1-2 contains *Hemiaulus incisus* Hajós, *Pyxilla gracilis* Tempére et Forti, *Trinacria excavata* Heiberg, *Rylandsia conniventa* Gleser, Dolmatova et Lupikina, *Coscinodiscus (Azpeitia) gombosii* Gleser, Dolmatova et Lupikina, *Kisseleviella cuspidata* Gleser, Dolmatova, et Lupikina, *Pinnularia* aff. *antiqua* Tscheremissinova, *Asterolampra vulgaris* Greville, *A. punctifera* (Grunow) Hanna, *Pseudotriceratium radiosoreticulatum* Grunow, *Thalassiosira dubiosa* Schrader, *Pseudopodosira corolla* (A. Schmidt) Hajós, *Sceptroneis vermiformis* Schrader, and others. The most characteristic species is *Rylandsia conniventa* Gleser, Dolmatova et Lupikina, which was found exactly in this sample and described for the first time (Gleser *et al.*, 1986). Representatives of the genus *Rylandsia* occur in the middle–upper Eocene deposits (Fenner, 1985; Barron and Baldauf, 1995). The *Rylandsia inequiradiata* Zone of upper Eocene–lower Oligocene was established in the South Ocean (Gombos and Ciesielski, 1983; Fenner, 1985). *Rylandsia biradiata* Greville is index species of the middle–upper Eocene beds (subzone "b") of the *Asterolampra marylandica* Zone in low latitudes (Fenner, 1984, 1985). Correlating these zones with the geomagnetic polarity scale of Cande and Kent (1992), Barron and Baldauf (1995) suggested their upper Eocene age.

In spite of presence of some Oligocene species (e.g., *Hemiaulus incisus* Hajós, *Pyxilla gracilis* Tempére et Forti, *Asterolampra punctifera* (Grunow) Hanna, *Pseudotriceratium radiosoreticulatum* Grunow, and others), Assemblage 4 should be attributed to the late Eocene, as it is lacking species characteristic of the lower Oligocene zonal assemblages from the northwestern Pacific (Oreshkina, 1996; Gladenkov and Barron, 1995; Gladenkov, 1998; Tsoy, 2002). Assemblage 4 can be conventionally defined as the *Rylandsia conniventa* assemblage, stratigraphic and geographic ranges of which need to be specified.

Silicoflagellates macerated from the same sample (Table 3) also evidence that Assemblage 4 is not younger than the late Eocene. They characterize the *Dictyocha hexacantha* Zone of the uppermost middle–lowermost upper Eocene in low and middle latitudes (Bukry and Foster, 1974; Bukry, 1977). These are *Dictyocha hexacantha* Schulz and *Corbisema ovalis* Perch-Nielsen, ranges of which are limited by this zone. Associated *Dictyocha deflandrei* Frenguelli ex Gleser, *D. frenguellii* Deflandre, and *D. pentagona* (Schulz) Bukry et Foster have their first occurrences in the zone, and *Dictyocha spinosa* (Deflandre) Gleser, *Bachmannocena paulschulzii* Bukry, and *Naviculopsis foliaceae* Deflandre disappear at the upper boundary of the zone (Bukry, 1981, 1984; Perch-Nielsen, 1985). Other species of wider stratigraphic range are nevertheless typical of the *Dictyocha hexacantha* Zone. The zonal index species is known from the middle Eocene Kellogg Shale, the middle–upper Eocene Kreyenhagen Formation of California (Barron *et al.*, 1984), and from the upper Eocene Oamaru Formation of New Zealand (Bukry, 1987).

Thus, silicoflagellates undoubtedly characterize the *Dictyocha hexacantha* Zone of the uppermost middle–lowermost upper Eocene. Since the diatom flora is represented by species typical of the same age interval, diatoms of the *Rylandsia conniventa* assemblage and silicoflagellates of the *Dictyocha hexacantha* Zone are supposed to correspond in age to the terminal middle-late Eocene.

The assemblage includes some diatom and silicoflagellate species of wider stratigraphic ranges, which are known from the Ommai Formation[1] and from the Telegraficheskii Cape Formation (lower reaches of the Anadyr River) of the upper Eocene (Sheshukova-Poretskaya, 1967; Nevretdinova, 1982). Some diatom species of the assemblage were also found in the Cape Tons Formation of the Karaginskii Island (Oreshkina, 1982, 1996), but they are associated with characteristic Oligocene forms.

It is to be noted that the middle Eocene interval in oceanic sediments of the Detroit Seamount not far away from the Kronotskii Bay (ODP Sites 883 and 884) is characterized predominantly by calcareous planktonic foraminifers and nannoplankton (*Proceedings...*, 1995; Basov, 1997). There are no diatoms in this interval, and rare radiolarians are represented by species typical of low latitudes, which are absent in the Kronotskii assemblages (Tsoy *et al.*, 2000).

CONCLUSIONS

Three diatom assemblages of middle Eocene age (those of *Lisitzinia kanayai, Lisitzinia inconspicua* var. *triloba*, and *Praecymatosira monomembranaceae* zones) and one presumably corresponding in age to the middle–late Eocene (diatoms of *Rylandsia conniventa* Zone) are established in volcanogenic–sedimentary deposits of the Kronotskii and Ol'ga canyons. The diverse silicoflagellate assemblage from these deposits is described for the first time and characterizes the *Dictyocha hexacantha* Zone. All the assemblages are supposed to have been formed mainly in bathyal environments and under a relatively high (almost subtropical) temperature of surface water.

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[1] The early Eocene planktonic and benthic foraminifers found in the Ommai Formation imply that it should be attributed to the lower–middle Eocene (*Resheniya...*, 1998).

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Correlation of the Upper Neopleistocene Alluvial and Lacustrine Sections, the Southern Trans-Urals Region

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Abstract—Composition and structure of the Upper Neopleistocene alluvial and lacustrine sections in the southern trans-Urals region are discussed. Sections of the Kamyshlov alluvial complex (Streletskoe, Vvedenskoe, Verkhnyaya Alabuga, Miasskoe, Bol'shoe Balandino) are correlated between each other based on their facies structure and with the lacustrine Kumlyak Formation from the Yuzhnyi section of the Baturino coal quarry. Paleontological remains (mammals, mollusks, and ostracodes) are characterized. Based on the comprehensive study of sections, correlation scheme of alluvial and lacustrine sediments is proposed.

Key words: Upper Neopleistocene, alluvium, lacustrine sediments, fauna, biostratigraphy, Kamyshlov Complex, Kumlyak Formation, paleogeography.

INTRODUCTION

Correlation of alluvial and lacustrine sediments is important for specification of the Quaternary stratigraphic scale in the southern trans-Urals region, for compilation of serial legends and maps of Quaternary sediments, and for reconstruction of natural environments of the Late Neopleistocene.

Investigation of fossil small mammals is one of the reliable methods to obtain information on burial conditions and age of sediments based on guide fossils. It is also important for reconstruction of sedimentation settings and for correlation with other regions.

Fossil remains of small mammals were found only in some outcrops of the Quaternary sediments during field works in the southern trans-Urals region although dozens of exposures in accumulative terraces were examined for this purpose. Finds of fossil microteriological remains in lacustrine sediments of interfluve areas, where Quaternary sediments were studied in industrial quarries and in many stratigraphic or prospecting boreholes, are less frequent.

In this work, we consider the results of study of the Upper Neopleistocene alluvial and lacustrine sediments in the southern trans-Urals region. We correlated sections of the Kamyshlov alluvial complex (Streletskoe, Vvedenskoe, Verkhnyaya Alabuga, Miasskoe, Bol'shoe Balandino localities) between each other, based on their facies structure, and with the lacustrine Kumlyak Formation exposed in the Yuzhnyi section of the Baturino coal quarry. All the sections are located close to 54–55° N (Fig. 1). We also reexamined collections of small mammals sampled in different years by A.V. Borodin, A.G. Maleeva, N. G. Smirnov, and V.V. Stefanovskii. All materials are stored at the Institute of Ecology of Plants and Animals, Uralian Division, Russian Academy of Sciences (Yekaterinburg), and at the Chair of Zoology of the Uralian State University (Yekaterinburg). Remains of freshwater mollusks and ostracodes are also used for correlation of alluvial and lacustrine sediments.

ALLUVIAL SEDIMENTS OF THE KAMYSHLOV COMPLEX

The *Kamyshlov terrace* is a most sustained geomorphologic element in all river valleys of the southern trans-Urals region. The terrace is of an accumulative type, 8 to 10 m high in general. Its surface is flat or

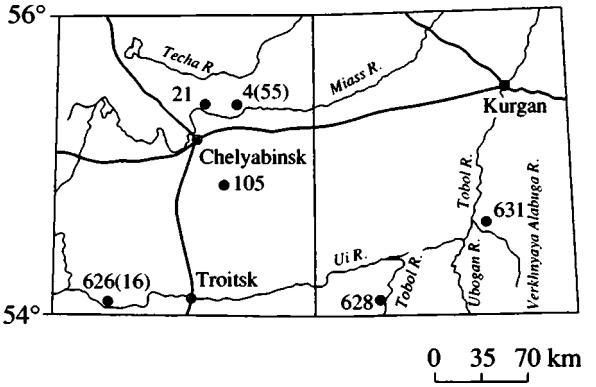

Fig. 1. Localities of reference exposures: (21) Bol'shoe Balandino, (4; 55) Miasskoe, (105) Yuzhnyi coal quarry, (626; 16) Streletskoe, (628) Vvedenskoe, (631) Verkhnyaya Alabuga.

slightly inclined riverward and, near the rear terrace margin, it can go up by 15 m because of a thick talus cover. The terrace is usually 1 to 2 km wide, being twice wider (3 to 5 km) in the lower courses. The following lithological-facies beds are distinguishable in the section of the Kamyshlov alluvial complex (from the base upward):

Bed 1. Talus and periglacial loams, brown, with columnar jointing, macroporous, calcareous. The lower boundary is usually complicated by cryogenic deformations (wedges, involutions). The thickness ranges from 2 to 5 m.

Bed 2. Periglacial (floodplain) silts or dusty fine- to medium-grained polymictic sands, brownish gray, with obscure wavy bedding, calcareous, with thin lenticular interbeds of silty clay or coarse-grained gravelly sand (roundness value 2 to 3). The lower boundary is erosional, wavy, with polymictic pebbles, ferruginate (intraformational hiatus). The thickness is 4–5 m.

Bed 3. Oxbow clays or lutite, lilac–gray, muddy, with plant detritus, frequently with cloddy or thin horizontal bedding, displaying ocherous ferruginous powdery coating on bedding surfaces, thin lentils and laminae of micaceous sand, and bearing rare shells of freshwater mollusks. The thickness is 1–2 m.

Bed 4. Fluvial shingle and gravely sands, gray, polymictic (roundness value 3 to 4), loose, commonly with oblique bedding, ferruginous and manganese spots, enclosing rare shells of freshwater mollusks and rodent remains. The sediments occur usually near to or below the water line. The thickness ranges from 2–3 to 5 m.

In the southern trans-Urals, there are known five localities of the Kamyshlov Alluvium.

The Verkhnyaya Alabuga locality (Exposure 631) is situated upstream of the eponymous village, 1.4 km away from the right tributary of the Verkhnyaya Alabuga River (Fig. 1). The section thickness near the rear suture of the terrace is 9 m. In this section, the Middle Neopleistocene compact clays of the Ui-Ubogan Formation rest on the erosional surface of dark gray siltstones with polymictic pebbles and gravel, which yield fossil remains of small mammals and freshwater mollusks (Bed 5). The siltstones are overlain by buried gley soil with small desiccation wedges (Bed 4) and by floodplain-periglacial sediments represented by polymictic sands with gravely lentils and thin interbeds, which yield remains of small mammals and shells of freshwater mollusks (Bed 3). The section is crowned by younger cross-bedded inequigranular loam (Bed 2) and contemporary soil (Fig. 2-I).

The Vvedenskoe locality (Exposure 628) is at the left steep bank of the Tobol River near the eponymous settlement (Fig. 1). The terrace is 8 to 10 m high here. It is composed of fluvial facies represented by cross-bedded inequigranular polymictic sands with remains of fossil rodents (Bed 7), which are intercalated with oxbow facies of bluish gray clayey silts slightly reduced in thickness (Bed 6) and with thicker periglacial dusty inequigranular sands (Beds 3–5). The alluvium is overlain by loamy cover of the Polar Urals time (Bed 2) and by soil (Fig. 2-II). Remains of fossil rodents are confined to cross-bedded sands of the fluvial facies (Bed 7).

The Streletskoe locality (Exposure 626; 16) is situated on the left steep bank of the Ui River near the eponymous settlement (Fig. 1). The section is composed here of all facies mentioned above. At the contact between periglacial and oxbow facies, there are large pseudomorphs after ice wedges and boulders of Paleozoic rocks (Fig. 2-III, Beds 3 and 4). The latter represent probably rock fragments brought and left by floating ice during spring floods. Fossils (remains of mammals, molluscan shells, ostracodes) were found in fluvial cross-bedded polymictic gravely sands (Bed 5) and in lenses of polymictic sands of the oxbow facies (Bed 4).

The Bol'shoe Balandino locality (Exposure 21) is discovered in the flux quarry near the eponymous village at the left bank of the Miass River (Fig. 1). In this locality, fluvial facies of polymictic shingle with inequigranular sandy matrix (Bed 9) rest on the karstified surface of Paleozoic marbles. Higher in the section, there are oxbow hydromica clays, dark to bluish gray, silty, enclosing sand nests, rare polymictic gravel, ferruginous spots, and shells of freshwater ostracodes (Bed. 8). The clays are, in turn, overlain by periglacial sediments represented by polymictic gravely sands with the clayey-loamy matrix, wavy bedding, marble blocks with crumpling signs underneath, clay lenses, ferruginous spots, and with bone remains of large mammals (Bed 7). The section is crowned by a member with horizontal-sinuous lamination (Bed 6) that is composed of alternating sandy hydromica clays and brownish gray clayey sands, inequigranular, locally ferruginous, having inclusions of marble debris and polymictic gravel (mostly quartz), and yielding fragments of molluscan shells and rodent remains. The alluvial section is overlain by a succession of the Nev'yansk buried soil of the Chernozem type (Bed 5), periglacial alluvium of the later accumulation cycle (Bed 2 and 3), talus cover (Bed 2), and recent soil (Fig. 3-I).

We examined sediment lithology throughout the section. The clayey components are hydromica, quartz grains, and mica flakes. The light fraction is composed of potassium feldspar (30 to 50%), quartz (25 to 45%), permanent mica (2 to 5%), and Fe hydroxides (up to 3%). In the heavy fraction, dominant are minerals poorly resistant to chemical weathering, such as hornblende (55 to 65%), epidote, and zoisite (10 to 15% together). Associated stable minerals are sphene (10 to 15%), ilmenite, leucoxene, and brookite (1 to 6%). The paleogeographic coefficient (ratio between stable and unstable minerals) is very low (0.07 to 0.14) that is characteristic of the Upper Neopleistocene sediments.

The Miasskoe locality corresponding to Exposure 4 (55) in the clay quarry is situated east of the eponymous

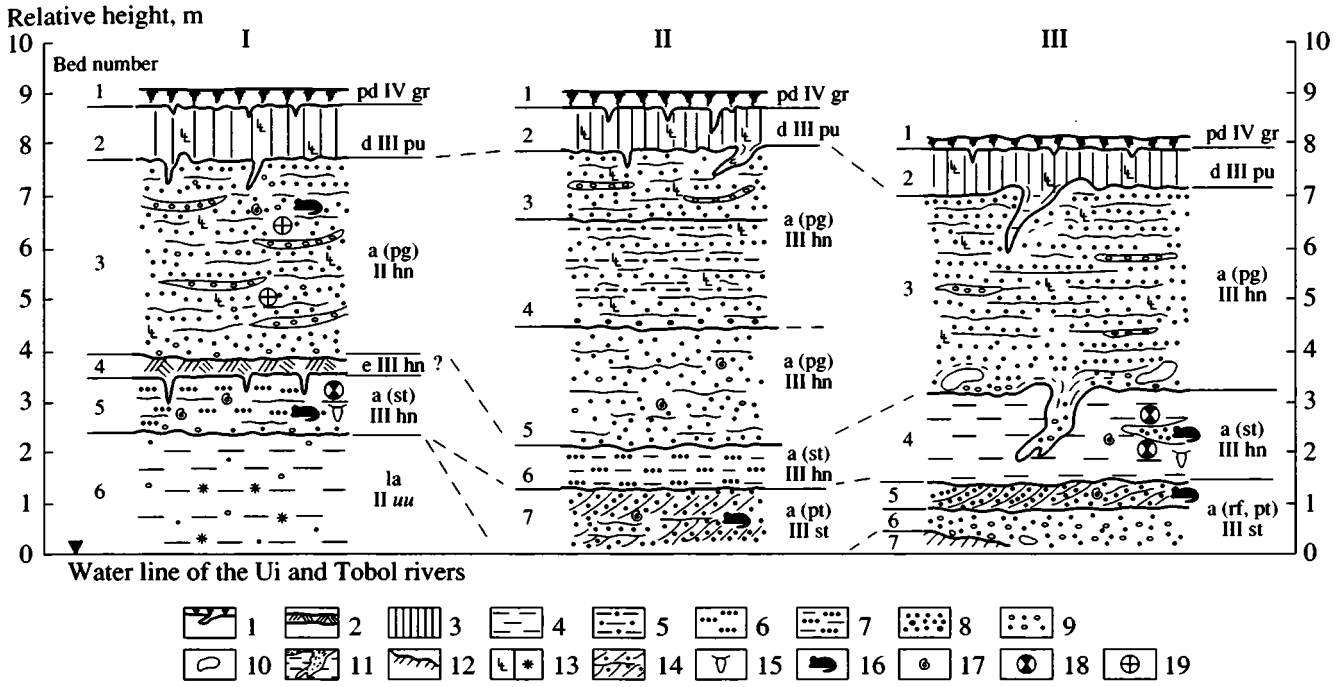

Fig. 2. Sketches of the Kamyshlov Alluvium sections in the Ui and Tobol river valleys, the southern trans-Urals region: (1) soil; (2) buried soil; (3) loam; (4) clay; (5) clayey sand; (6) silt; (7) clayey silt; (8) inequigranular sands; (9) gravely shingle with sandy filler; (10) boulders; (11) pseudomorphs after ice veins; (12) Paleozoic rocks (socle); (13) calcified and ferruginate rocks; (14) bedding type; (15) large mammal, (16) rodent, (17) molluscan, and (18) ostracode remains; (19) spores and pollen. Sections: (I) Exposure 631, right bank of the Tobol River upstream of the Verkhnyaya Alabuga River, 1.4 km of the synonymous village; (II) Exposure 628, left bank of the Tobol River near the Vvedenskoe Settlement; (III) Exposure 626 (16), left bank of the Ui River near the Streletskoe Settlement. Regional stratigraphic units: (Ivgr) Holocene Gorbunov Horizon; upper Neopleistocene: (IIIpu) Polar Urals (Ostashkovo) Horizon, (IIInv) Nev'yansk (Leningrad) Horizon, (IIIhn) Khanmei (Podporozhskii) Horizon, (IIIst) Streletskoe (Mikulinian) Horizon; lower Neopleistocene: (Iiuu) Ui-Ubogan Formation of the Middle Urals Horizon, (P_2 sr) Eocene Serov Formation; (P_2 ir) Eocene Irbit Formation; (PZ) Paleozoic. Indices of sediment genetic types: (a) fluvial, (d) talus, (la) lacustrine-fluvial, (l) lacustrine, (e) eluvial (buried soil), (pd) soil-peat; (rf) deposited, (pt) potamoclastic, (st) oxbow, and (pg) periglacial fluvial facies.

village in the Miass River basin (Fig. 1). Polymictic gravely sands of the fluvial facies (Bed 7) are overlain here by oxbow clays (Bed 6) with shells of freshwater mollusks. The overlying periglacial alluvium (Bed 5) is composed of brownish to brown calcareous sands locally displaying horizontal-wavy lamination emphasized by thin (up to 1 cm) lentils of gravely sand or by black accumulations of dark-colored minerals and enclosing remains of large and small mammals (Fig. 3-II). Similarly to the Bol'shoe Balandino section, the Kamyshlov Alluvium is overlain here by the Nev'yansk buried soil (Bed 4), periglacial sediments (Bed 3), and talus loam of the Polar Urals time (Bed 2).

FOSSILS FROM THE KAMYSHLOV ALLUVIUM

Flora. The spore-pollen spectra (102 to 138 grains) from silty clays (Bed 8) of the Bol'shoe Balandino locality (Exposure 21) show the prevalence of herbaceous pollen (98.2%) throughout the bed (6.4 to 7.6 m), in which mesophytic Caryophyllaceae, Compositae, and *Polygonum bistorta* L. are dominants (71.5 to 76.1%). Upper spectra from the interval of 6.3 to 6.6 m show presence of Cruciferae and Umbelliferae pollen. Xerophytes (23.9 to 28.5%) are represented by *Artemisia* sp., Chenopodiaceae, and Gramineae pollen, the latter characteristic of the lower part (7.2 to 7.6 m). Single grains of arboreal pollen represent *Pinus sylvestris* L., *Betula pubescens* Ehrh., and *Salix* sp. The spectra are remarkably uniform, and only those from the upper part of the show a higher diversity of mesophytic taxa. The spectra characterize phytocoenoses of the open meadow-steppe landscapes of the Khanmei (Valdayan) cooling. Similar, although discrete spectra with prevailing xerophytes (Chenopodiaceae-*Artemisia*) groupings) are recorded in the periglacial alluvium (Bed 3) of the Verkhnyaya Alabuga section (Exposure 631).

Mollusks. Molluscan remains occur in all examined sections of the Kamyshlov Alluvium (Table 1). Prevalent in the molluscan faunas of the southern trans-Urals region are transit species characteristic of the Neopleistocene sediments. Moreover, species *Lymnaea bolotensis* and *L.* cf. *ventricosella* appear in the Late Neopleistocene only. The molluscan assemblages mainly consist of the European-West Siberian species associated with less abundant European, Eurasian, and Siberian forms.

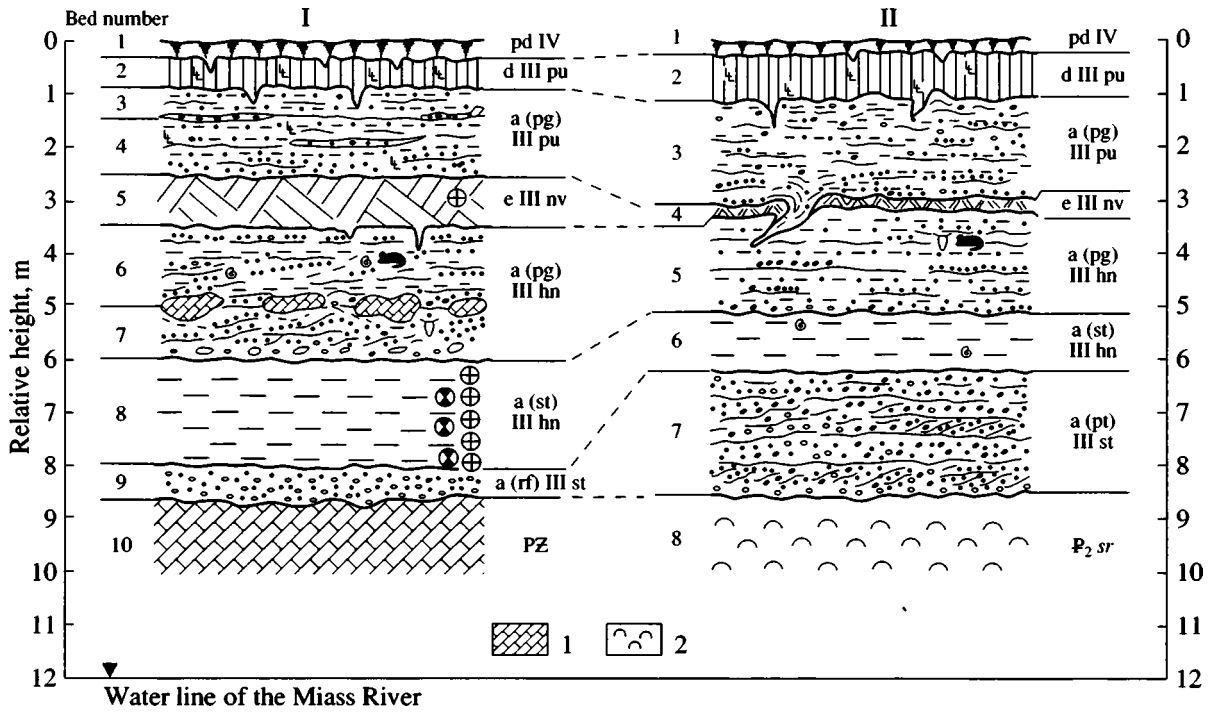

Fig. 3. Sketches of the Kamyshlov Alluvium structure in the Miass River valley of the southern trans-Urals region: (1) marble; (2) opoka (other symbols as in Fig. 2). Sections: (I) Exposure 21, left bank of the Miass River, flux quarry near the Bol'shoe Balandino Village; (II) Exposure 4(55), western wall of the brick quarry in the eastern outskirts of the village of Miasskoe at the left bank of the Miass River.

Freshwater mollusks are mainly represented by eurythermal dwellers of accessory water basins and by subordinate cryophilic forms *Valvata sibirica, Anisus leucostoma, Euglesa obtusalis*, and *Lymnaea peregra.* Most abundant molluscan remains were found in oxbow clays of the Verkhnyaya Alabuga (Bed 5) and Streletskoe (Bed 5) localities. Six species in common characterize both localities. The peculiar feature of the Verkhnyaya Alabuga molluscan fauna is presence of European and absence of European-Siberian species, though the latter are registered in the Streletskoe locality (Table 1).

In the Vvedenskoe (Bed 5), Bol'shoe Balandino (Bed 6), and Miasskoe (Bed 5) localities, molluscan remains were sampled from periglacial alluvium overlying the oxbow facies. Freshwater mollusks from these localities are represented by species that are present in the cryophilic molluscan assemblage from loess sections of the Russian Plain (Danilovskii, 1955).

It is likely that number of cold-resistant molluscan species increases from older to younger Neopleistocene sediments (Table 1). These are one of 10 species (n = 334) in the Streletskoe locality (Bed 5), 3 of 18 species (n = 338) in the Verkhnyaya Alabuga locality (Bed 5), 2 of 7 species (n = 65) in the Bol'shoe Balandino locality (Bed 6), and 3 of 7 species (n = 21) in the Miasskoe locality (Bed 6).

Ostracodes. Fossil crustaceans were found in oxbow clays of the Verkhnyaya Alabuga (Exposure 631), Streletskoe (Exposure 626), and Bol'shoe Balandino (Exposure 21) sections. Their assemblages differ in species composition: *Cyclocypris*, *Candona*, and *Candoniella* species prevail in the Verkhnyaya Alabuga and Streletskoe sections, whereas *Limnocythere* forms are dominant in the Bol'shoe Balandino locality (Table 2). Some researchers (Arkhipov *et al.*, 1973) consider the decreasing proportions of *Limnocythere* forms in ostracode fauna as a result of shoaling, desiccation, and, probably, freezing of water basins. The fossil ostracode assemblage contains both eurythermal and cryophilic species (*C. neglecta, C. rectangulata*). Preponderant are species widespread in the Middle-Upper Neopleistocene sediments.

Large mammals. Remains of large mammals usually represented by poorly preserved bone fragments occur as single specimens.

In the Verkhnyaya Alabuga locality, the bed of dark gray siltstone (Bed 5 of Exposure 631) yielded bone fragments of *Canis lupus* L., (small specimen), *Mammuthus* sp., *Coelodonta antiquitatis* Blum., *Bison* sp., and *Saiga tatarica* L. (determinations by I.E. Kuz'mina, Zoological Institute RAS). The bed of oxbow clays (Bed 4) in Exposure 626 (16) of the Streletskoe locality contains fragments of the right radial and left humeral bones belonging to *Equus* ex gr. *caballus* L. (determi-

Table 1. Mollusks from alluvial and lacustrine sediments of the southern trans-Urals (determinations by L.I. Krylova, Uralian Teacher's Training College, Yekaterinburg)

Species \ Locality	Alluvium					Lacustrine sediments	
	Verkhnyaya Alabuga, Bed 5	Streletskoe, Bed 5	Vvedenskoe, Bed 5	Miasskoe, Bed 5	Bol'shoe Balandino, Bed 6	Yuzhnyi, Bed 3	Yuzhnyi, Bed 5
Terrestrial							
Pupilla muscorum L., 1758				4			2
Succinella oblonga Drap.					5		
Succinea putris L., 1758	16	25		1			
Vallonia excentrica Sterki				11			
European-West Siberian group						4	
Lymnaea truncatula, Müll., 1774	2						
L. peregra Müll., 1774	23	2					25
L. palustris Müll., 1774					7		
Bithynia troscheli Paasch, 1842		18					5
B. inflata Hans., 1845	80						
B. tentaculata L., 1758			1				
Planorbis planorbis L., 1758	18	75		1	10		26
P. corneus L., 1758	4	1					
Anisus leucostoma Millet, 1813	3	183			11		6
A. vortex L., 1758	8					1	
Choanomphalus rossmaessleri Schmidt, 1851	20					9	1
Euglesa hibernica West., 1894							
E. obtusalis Pfeiff., 1821	6						
Eurasian group							
Lymnaea stagnalis L., 1758		2		1			36
L. saridalensis Mozley							1
Amesoda asiatica Mart			3				
Musculium creplini Dunk., 1843	15						
Physa fontinalis L., 1758	6						
European group						1	
Euglesa casertana Poli, 1791	41						6
Anisus strauchianus Cles., 1886	4						
Amesoda draparnaldi Cles., 1879					1		
Choanomphalus stelmachoetius Bourg., 1860	5						
European-Siberian group						1	
Euglesa nitida Jen., 1832							4
Anisus spirorbis L., 1758		3					
Anisus acronicus Fer., 1807					6		
Armiger crista L., 1758				1			
Siberian group							
Lymnaea ventricosella Dyb.		31					9
L. bolotensis Mozley	13	4			25		
Valvata alpestris Kust.	24						
V. sibirica Midd., 1851	50					2	
Valvata sp.	1		1				
Euglesa sp.				2			
Lymnaea sp.			1	1			

Table 2. Ostracodes from alluvial and lacustrine sediments of the southern trans-Urals (determinations by E.D. Yaskevich* and S.M. Vitskikh,** Open Joint-Stock Company UGSE)

Species \ Locality	Alluvium			Lacustrine sediments		
	Bol'shoe Balandino, Bed 8*	Streletskoe, Bed 4*	Verkhnyaya Alabuga, Bed 5*	Yuzhnyi, Bed 5**	Yuzhnyi, Bed 4**	Yuzhnyi, Bed 3**
Ilyocypris bradyi Sars.		+	+	11	24	67
I. curtus Jask.				14	1	6
I. getica Mazi.					5	24
I. bella Schar.						3
I. gibba Ramd.		+			5	11
I. caspiensis Neg.	6					2
Cyclocypris longus Neg.		+	+		6	5
C. triangulata Neg.				1		5
C. laevis Müll.		+	+	22	49	5
C. globosa Sars.		+	+		7	1
Candona candida Müll.					2	8
C. rostrata Br. et Norm.		+			11	10
C. neglecta Sars.		+	+		11	12
C. arcina Liep.			+	6	15	30
C. stagnalis Sars.						2
C. rectangulata Alm.		+			1	1
C. fabaeformis Fisch.					3	11
C. sarsi Hart			+			3
Eucypris grassa Müll.			+			
Candoniella subellipsoida Schar.	4	+		46	156	170
C. albicans Brady	2	+		5	17	54
C. lunata Jask.		+		5		19
C. schubinae Mand.			+	3		
Cyprideis littoralis Br.			+		4	
C. torosa Jones			+		2	
Cypria deplanata Jask.		+	+			
Cypris aff. *pubera* Müll.		+				
Limnocythere vara Liep.					1	16
L. dorsotuberculata Neg.	17		+	13	27	136
L. inopinata Baird.	1			9	4	74
L. aperta Jask.	1					4
L. parallela Dieb						1
L. manytschensis Neg.				2	12	35
L. pseudoconcava Neg.						1
L. postconcava Neg.	1		+		25	19
L. postconcava var. *tuberculata* Neg.						7
L. certa Jask.					1	3
L. dorsoelovata Jask.						4
L. stapliny Gut. et Bens.						7
L. grinfeldi Liep.	1					1
L. originalis Neg.		+				
Darwinula stevensoni Br. et Rob.					3	

nation by P.A, Kosintsev, Institute of Ecology of Plants and Animals, Uralian Division RAS). Periglacial alluvium (Bed 5) from Exposure 4 (55) of the Miasskoe locality yielded metatarsal and radial bones of *Equus caballus* subsp., and also femur fragment and epistropheus of *Bison* sp. (determinations by E.A. Vangengeim, Geological Institute RAS).

All these remains belong to animals of the mammoth assemblage.

Small mammals. Most abundant remains of small mammals about (1550 fragments of 29 species, Table 3) were found in the Verkhnyaya Alabuga locality (Exposure 631). The fauna is transitional from the steppe to desert-steppe type. Voles are represented by dominant *Lagurus* cf. *lagurus* and *Eolagurus* cf. *luteus*. Abundant amid remains are bones of *Ochona* cf. *pusilla, Marmota* cf. *bobac, Spermophilus* ex gr. *pygmaeus, Allactaga* ex gr. *jaculus*, and *Allactagulus* sp. It is remarkable that the assemblage lacks voles of the genus *Clethrionomys* and *Microtus oeconomus*. The former are typical of the forest zone and dwell in floodplain shrub thickets of the tundra and steppe zones. Root voles are distributed in both the moistened meadow-type habitats and shrub thickets that shelter their burrows (Gromov and Polyakov, 1997; Ognev, 1950). Consequently, the shrub vegetation during the period of host sediment accumulation was probably very scarce even in coastal areas. Another species *Arvicola terrestris* that is also present in the assemblage under consideration is typical of near-water biotopes can occupy the coastal sedge-reed thickets of water basins, when shrub vegetation is missing (Ognev, 1950).

The fauna can be dated and correlated with others based on its species composition and morphological characteristics of voles. The presence of *Arvicola terrestris* suggests that the fauna under consideration is correlative with that of the MQR 1 Zone of East Europe (Vangengeim *et al.*, 2001), thus being of the Late Neopleistocene or younger age. Based on dimensions of the lower buccal tooth and enamel differentiation degree, Maleeva and El'kin (1986) attributed the water vole remains to subspecies *Arvicola terrestris alabugensis* transitional between the Middle Neopleistocene and recent forms.

Lagurus lagurus from the Verkhnyaya Alabuga locality was larger (Table 4) than representatives of this species from the Gadyach locality (terminal Middle Neopleistocene, Markova, 1982) and smaller than the subspecies *Lagurus lagurus major* described from the upper Valdayan sediments (Rekovets, 1985). The comparison of teeth morphotype characteristics of fossil lemmings from the Verkhnyaya Alabuga locality with those of recent subspecies, using methods of the multidimensional analysis, showed that they can be dated with confidence back to the first half of the Late Neopleistocene.

When describing morphological characteristics (teeth dimensions, conjunction degree of triangles) of *Eolagurus luteus*, Maleeva (1982) noted that the yellow steppe lemming from the Verkhnyaya Alabuga locality is transitional between the Middle Neopleistocene *Eolagurus luteus volgensis* and recent *Eolagurus luteus*.

The narrow-skulled vole (*Microtus gregalis*) from the Verkhnyaya Alabuga locality is relatively large and shows significant variability in the length of the chewing surface in the first lower molar (Table 5). The species is, like the Neopleistocene *M. gregalis* from West Siberia, smaller than its present-day subspecies dwelling in tundra and larger than the steppe subspecies. (Smirnov *et al.*, 1986). The evolutionary level (age) of *M. gregalis* populations from different localities is determined by the extent of angle development on the buccal side of the unpaired anterior loop. The absence of the angle characterizes the most primitive gregaloid morphotypes. Judging from proportions of these morphotypes (Table 5), the Verkhnyaya Alabuga assemblage is more advanced than that from the Yarsino locality (terminal Middle Neopleistocene), but more primitive as compared with the assemblage from the locality of 430th km that is 24 ka old (Smirnov *et al.*, 1984). Thus, *Microtus gregalis* from the Verkhnyaya Alabuga locality is younger than the Middle Neopleistocene and not older that the Valdayan epoch. The comparison of the Verkhnyaya Alabuga assemblage with the Pleistocene populations of narrow-skulled voles in East Europe shows that the former is similar to the early Valdayan assemblage from the Gralevo-2 locality (Motuzko, 1992). In both assemblages, the proportion of primitive morphotypes exceeds 50%, but *Microtus gregalis* from the Gralevo-2 locality is smaller than the same species from the Verkhnyaya Alabuga assemblage (dimension parameters 2.3–2.61–3.0 and 2.25–2.72–3.2, respectively), though this can be regarded as a result of geographic variability.

Based on species composition of small mammals and morphologic characteristics of molar teeth of voles, the Verkhnyaya Alabuga assemblage can be dated back to the first half of the Late Neopleistocene (Khanmei Horizon).

The vole assemblage from the Vvedenskoe locality is dominated by *Eolagurus luteus* and *Lagurus lagurus* associated with *Allactaga* species, *Cricetulus* ex gr. *migratorius*, and *Ochotona* cf. *pusilla* (Table 3.) Such a composition may indicate the similarity of habitat environments around this and Verkhnyaya Alabuga localities. However, some peculiar features are characteristic of the assemblage under consideration. Par example, a Chiroptera skull was found in the Vvedenskoe section that is the only one in the trans-Urals with remains of these mammals whose distribution areas are limited by availability of necessary refuges. In steppe and semi-desert biotopes, they populate hollows of old trees in floodplains (Kuzyakin, 1950). Another peculiarity of the Vvedenskoe assemblage is the absence of narrow-skulled and presence of root voles, which populate

Table 3. Small mammals from alluvial and lacustrine sediments of the southern trans-Urals region (determinations by L.I. Krylova, Uralian Teacher's Training College, Yekaterinburg)

Taxon \ Locality	Alluvium						Lacustrine sediments		
	Vvedenskoe, Bed 7	Streletskoe, Bed 5	Streletskoe, Bed 4	Verkhnyaya Alabuga, Bed 5	Miasskoe, Bed 5	Bol'shoe Balandino, Bed 6	Yuzhnyi, Bed 5	Yuzhnyi, Bed 4	Yuzhnyi, Bed 3
Insectivora									
Erinaceus sp.									4
Desmana moschata L., 1758								1	1
Sorex ex gr. *araneus* L., 1758								1	
Sorex sp.				1					1
Chiroptera	1								
Lagomorpha									
Lepus sp.		1		2					
Ochotona cf. *pusilla* Pall., 1768	1			72					4
Rodentia									
Spermophilus sp.		2					4	4	1
Spermophilus ex gr. *pygmaeus* Pall., 1778				26					
Marmota cf. *bobac* Müll., 1776				29					
Allactaga ex gr. *jaculus* Pall., 1778	1			9					
Allactaga sp.	1								
Allactagulus sp.				5					
Ellobius ex gr. *tancrei* Blas., 1884				14					
Cricetulus sp.								1	
Cricetulus ex gr. *migratorius* Pall., 1773	1			5					
Clethrionomys sp.		1							
Lagurus cf. *lagurus* Pall., 1773	4	7	5	751	15			28	14
Eolagurus cf. *luteus* Eversm., 1840	10	2	2	483	1	1		13	8
Lemmus sp.							1	1	
Arvicola terrestris alabugensis Maleeva et Elkin, 1986				76					
Arvicola terrestris L.,1758			4				4	9	8
Microtus gregalis Pall., 1779		2	2	70			1	7	2
Microtus oeconomus Pall., 1776	1	2					1	7	
Microtus ex gr. *arvalis* Pall., 1779									1
Microtus sp.	1	13	3			2	1	37	24
Total	21	30	16	1541	16	3	12	109	67

moistened open meadow-type biotopes immediately near shrub thickets. These peculiarities characterize the past semidesert-steppe biotope with shrub thickets and, probably, with trees in floodplains.

According to large dimensions, the steppe lemming (*L. lagurus*) from the Vvedenskoe section is similar to the counterpart from the Ozernoe locality of East Europe (Table 4) dated back to the terminal Middle Pleistocene (Rekovets, 1994). The yellow steppe lemming is also characterized by large dimensions (the total length of chewing surface is 3.45). In the conjunction degree of triangles and dimensions, this form is close to *Eolagurus luteus volgensis* Alexandr. described from the Middle Neopleistocene.

The species composition of small mammals from the Vvedenskoe locality, morphological characteristics of vole molars, and position of host beds within the section imply that stratigraphic range of the assemblage is

Table 4. Size parameters of the first lower molars of steppe lemmings

Locality	*n*	Length	Width	Length of protoconid
Vvedenskoe	2	2.65; 2.7	1.0; 1.0	1.525; 1.575
Verkhnyaya Alabuga	324	2.2–2.57–3.0	0.65–0.91–1.1	1.25–1.51–1.75
Streletskoe, Bed 4	1	2.45	0.825	1.425
Yuzhnyi, Bed 4	6	2.2–2.45–2.625	0.85–0.89–1.00	1.15–1.325–1.45
Yuzhnyi, Bed 3	1	2.6	1.0	1.45
Miasskoe, Bed 5	1	2.6	1.0	1.45
Ozernoe*	10	2.6–2.7–2.8	0.8–0.89–0.95	1.3–1.47–1.65
Gadyach**	3	2.1–2.4–2.7	0.85–0.9–0.95	
Novgorod-Severskii***	50	2.44–2.66–3.0	0.8–0.96–1.1	1.2–1.45–1.6
Arapovichi**	13	2.35–2.5–2.65	0.9–0.94–1.15	

* After Rekovets, 1994, ** after Markova, 1982, *** after Rekovets, 1985.

correlative with both the MQR 2 and MQR 1 zones of the East European scale (Vangengeim *et al.*, 2001). It may correspond to the Streletskoe (Mikulinian) interglacial interval of the Late Neopleistocene.

In Exposure 626 (16) of the Streletskoe locality, remains of small mammals were found at two levels: in fluvial cross-bedded sands (Bed 5) and in lenticular intercalations of polymictic sands in clayey oxbow facies (Bed 4). The assemblage of Bed 5 is dominated by voles *Microtus gregalis, M. oeconomus*, and *Lagurus* sp. found in association with *Lepus, Spermophilus*, and *Clethrionomys* (Table 3). The assemblage characterizes a steppe biotope with shrub thickets in floodplains of rivers. Lenses of polymictic sands in clayey oxbow facies (Bed 4) yield remains of voles, prevalent among which were *L. lagurus* and *Arvicola terrestris.* Remains of *E. luteus* and *M. gregalis* occur in equal proportions. In distinction from the previous lower assemblage, this one lacks species representing dwellers of shrub thickets. Based on presence of *Arvicola terrestris*, the Streletskoe fauna as whole can be correlated with that from the East European zone MQR 1 of the Late Neopleistocene or younger age (Vangengeim *et al.*, 2001).

Guide vole forms from the locality under consideration are scarce, inappropriate for the adequate analysis of their succession in the section. It should be noted, however, that both the steppe and yellow lemmings are characterized by small dimensions (Table 4). This feature is also peculiar of steppe lemmings from the Arapovichi locality of the Russian Plain (Markova, 1982), where it is dated back to the first half of the Late Neopleistocene. Based on species composition and morphological characteristics of lemmings' molars, the assemblage of small mammals from the Streletskoe locality can be referred to the lower half of the Upper Neopleistocene.

Two of three determinable molar teeth from the Bol'shoe Balandino locality (Exposure 21) belong to *Microtus* sp. and one to *Eolagurus* sp. (Table 3). These genera are characteristic of steppe and semidesert landscapes, although scarcity of their remains and absence of teeth enabling species identification do not allow us to characterize in detail the biotope and correlation with other faunas.

Rodent remains from Exposure 4 (55) of the Miasskoe locality (Bed 5) represent species of the Lagurini tribe only, such as *Lagurus lagurus* (15 specimens) and *Eolagurus* cf. *luteus* (1 specimen). These species are typical of steppe and semidesert landscapes, and we cannot add anything more definite because of scarcity of their remains. The steppe lemming from the Miasskoe locality is larger than its counterpart from the Verkhnyaya Alabuga section, but smaller than that from the Novgorod-Severskii locality (Table 4). In our opinion, the assemblage from this locality is older than fauna from the Novgorod-Severskii section but younger than the Verkhnyaya Alabuga assemblage, thus being correlative with the uppermost part of the Khanmei Horizon.

In general, the small mammal fauna from all examined localities of the Kamyshlov Alluvium is represented mostly by voles of the *Lagurus, Eolagurus*, and *Microtus* genera, all indicative of steppe and semidesert

Table 5. Morphological characteristics of the first lower molars of narrow-skulled voles (n is number of specimens)

Locality	*n*	Length of M_1	The proportion of the gregaloid morphotype, %
Verkhnyaya Alabuga	50	2.25–2.72–3.2	31
Arsino	79	2.4–2.67–3.00	38
430 km	76	2.33–2.63–3.00	28
Gralevo-2*	57	2.3–2.61–3.0	35

* After Motuzko, 1992.

environments of their habitats. This inference is consistent with presence of bone fragments representing genera *Spermophilus, Marmota, Allactaga*, and *Cricetulus* in the studied assemblages.

LACUSTRINE SEDIMENTS OF THE KUMLYAK FORMATION

The southern trans-Urals is a region of widespread lacustrine sediments. These deposits fill in the ancient lakes and depressions, which are manifested in the present-day relief as areas of insignificant lowering and bogs.

One of the reference sections characterizing these sediments was examined in the Yuzhnyi coal quarry 2.5 km southward of the town of Emanzhelinsk (Fig. 1). In the central part of the northern quarry wall, there is exposed the following succession of beds (from the top downward):

0.0–0.2 m 1. Soil.

0.2–0.6 m pgIIIpu 2. Light brown sandy loam, macroporous, calcareous, with columnar jointing; the lower boundary is gradual, locally with small ice wedges (a contemporaneous freezing zone).

0.6–4.5 m lIIIhn 3. Greenish gay clayey silt, calcareous (up to extent of clayey marl), moderately compact, with weak horizontal bedding, quartz gravel (roundness value 3 to 4), ferruginous spots, fragments of molluscan shells, and mammal remains.

4.5–6.3 m lIIIhn 4. Wavy alternation of gray to greenish gray beds (2–3 to 7–10 cm thick) composed of silt, clayey silt, and sand-size clay rolls; rocks are calcareous (locally marly), with rare siliceous-quartzose pebbles, gravel (roundness value 3 to 5 points), rounded clasts of limy concretions, molluscan shells, and mammal remains. The lower boundary is gradual.

6.3–8.0 m a(pt) IIIst 5. Dirty gray polymictic inequigranular sands with wavy-cross bedding and silty–clayey matrix; rocks are calcareous, enclosing quartzose and siliceous gravel and pebbles, carbonate concretions (roundness value 3 to 5), molluscan shells, and rodent remains. The lower boundary is distinct, erosional.

8.0–9.0 m P_2ir 6. Light gray weathered diatomites (the Eocene Irbit Formation).

The basal facies in this section (Bed 5) represents probably alluvium of small steppe rivers that is overlain by lacustrine sediments (Beds 4, 3). The alluvium lower boundary with diatomites is erosional, while lacustrine sediments show gradual transition to alluvium and have cryogenic boundary with diatomites, which capriciously outlines injection-like cryogenic wedges and tongues (Fig. 4). Glacial deformations presumably developed in two stages. During the Leplino (Moscovian) time, prior to accumulation of alluvial sediments (Bed 5), the surface of weathered Eocene diatomites (Irbit Formation) was deformed under influence of "glacial diapirism" with formation of small U-shaped channels of the thermokarst type, along which the intermittent streams flowed. The Khanmei (Valdayan) or Polar Urals (Ostashkovo) glaciations resulted in intense development of permafrost that led to deformation of the glacial structure walls and to formation of injection-like wedges and subhorizontal tongues, the pseudomorphs after ice veins.

FOSSILS FROM LACUSTRINE SEDIMENTS OF THE KUMLYAK FORMATION

Flora. Spore–pollen spectra turned out to be representative only for alternating silts (Bed 4, depth interval of 4.9–5.8 m) and greenish gray clayey silts (Bed 3, depth interval of 3.8–4.2 m), although samples for the palynological analysis were collected throughout the entire Kumlyak Formation section.

Spectra characterizing Bed 4 are of the forest–steppe type, consisting of arboreal (44.1 to 56.3%) and herbaceous (42.5 to 54.2%) pollen and spores (1.2 to 2.9%). The arboreal pollen is dominated by that of *Pinus sylvestris* (68.5 to 85.5% of the group in total), and subordinate components represent *Abies sibirica* L. (up to 2%), *Pinus sibirica* Rugh., *Pinus* sp., *Salix* sp., *Betula pubescens* Echr. (up to 2%), *B. humilis* Schranc (up to 4%), *B. ventricosa* Echr. (up to 3%), and *Betula* sp. (3.8 to 13.7%). The herbaceous pollen is of the following composition: xerophilous Chenopodiaceae (43.8 to 54.1%) and *Artemisia* sp. (21.9 to 35.1%), mesophilic Compositae (16.2 to 34.4%), *Polygonum bistorta* L. (up to 8.1%), Rosaceae, and Campanulaceae (up to 2.7%). Spores of Polypodiaceae and *Sphagnum* sp. occur as single grains.

In the lower part of greenish gray clayey silts (Bed 3), spore-pollen spectra are of the steppe type: arboreal 1.2 to 3.2%, herbaceous 95.2 to 96.8%, and spores 1.6 to 2.1%. Arboreal pollen is represented by single grains of *Picea* sect., *Eupicea, Pinus sylvestris* L. Prevalent amid xerophytes are *Artemisia* sp. (58.6 to 60.2%) and Chenopodiaceae (20.4 to 22.5%). Subordinate pollen of meadow herbaceous includes representatives of Compositae (8.3 to 12.9%), Gramineae (3.3 to 5.8%), Cyperaceae (up to 1.6%), Caryophyllaceae (1.1 to 2.4%), Cruciferae, Umbelliferae (0.8 to 1.1%). Present as single grains are Polypodiaceae spores.

Spore–pollen spectra from laminated silts (Bed 4) characterize landscapes of light coniferous forests with meadow–steppe areas in flat interfluves. They are correlative with a transitional climatic phase that separated the glacial-humid and glacial-arid epochs of the Khanmei (Valdai) glaciation. Spectra with dominant xerophytes from the upper silts (Bed 3) characterize periglacial steppe landscapes corresponding to the culmination of the glacial-arid phase of the Khanmei glaciation.

Mollusks. The molluscan assemblage consists mostly of East European–West Siberian species (50 and 60% in Beds 5 and 3, respectively) accompanied by European, European–Siberian, and Eurasian forms

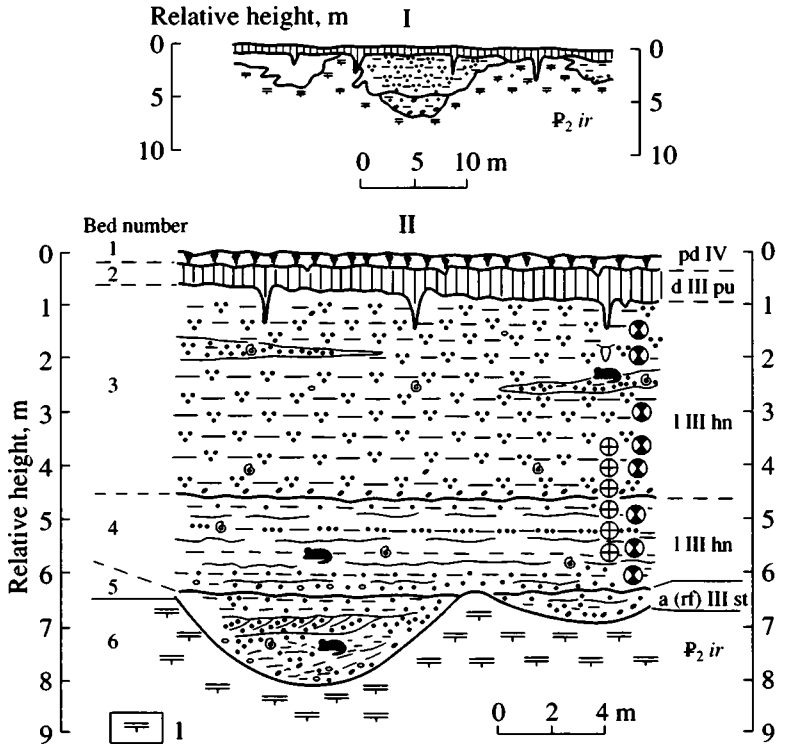

Fig. 4. Sketch of the lacustrine sedimentary section in the northern wall of the Yuzhnyi coal quarry, the southern trans-Urals region: (1) diatomite (other symbols as in Fig. 2). Sections: (I) Exposure 105, general view; (II) Exposure 105, central part.

(Table 1). Most of them are dwellers of both accessory and floodplain water basins. *Euglesa nitida*, which occur nowadays only in floodplain water basins (Dolgin, 2001), can serve as an indicator of lacustrine settings. The freshwater molluscan assemblage is dominated by eurythermal species associated with subordinate cold-resistant forms (*Anisus leucostoma* and *Euglesa nitida*).

Ostracodes. The abundant and diverse ostracode assemblage from the Kumlyak Formation consists of species representing genera *Ilyocypris, Cyclocypris, Candona, Candoniella*, and *Limnocythere*. Most of them are transit species occurring throughout the entire Neopleistocene section (Table 2). Many *Limnocythere* and *Ilyocypris* species, such as *L. manytschensis, L. stapliny, L certa, L. aperta, I. curtus, I. caspiensis*, and *I. bella*, become extinct in the Neopleistocene and do not occur in younger sediments. Prevalent amid ostracodes are eurythermal and eurygamous crenophilous forms. The share of cold-resistant species *Candona candida, C. neglecta, C. stagnalis*, and *C. rectangulata* does not exceed 4% of the total assemblage.

Large mammals. The silt member of lacustrine origin (Bed 3) encloses bones of large mammals of the mammoth assemblage consisting of *Capreolus capreolus* L., *Bison* sp. (determination by P.A. Kosintsev, Institute of Ecology of Plants and Animals, Uralian Division RAS), *Mammuthus primigenius* Blum. (young type) and *Coelodonta antiquitatis* Blum. (determinations by E.A. Vangengeim, Geological Institute RAS).

Small mammals. Remains of small mammals were washed out from polymictic sands (Bed 5), alternating silts and sands (Bed 4), and from a sand lentil within clayey silt (Bed 3).

Although mammal remains are scarce in Bed 5 (Table 3), they represented mostly species typical of or common in open habitats (Borodin, 1997), thus enabling interpretation of past landscapes and biotopes. It is remarkable that of the only steppe species (*Spermophilus* sp.) is present in the assemblage. At the same time, we found a molar of *Lemmus* sp. characteristic of tundra environments. The joint occurrence of gopher and lemming remains used to be considered as disharmonic and "lacking analogues." It should be noted however that present-day lemmings populate also the boggy thin-forested areas of the forest zone (Ognev, 1950), where they prefer substantially moistened areas covered with moss, sedge, and cotton-grass. Gophers can sporadically appear in the forest zone as well (Marvin, 1969). Based on these observations, we assume that the region around the accumulation site of sediments under consideration included moistened areas of thin forests with developed moss cover and drained slopes suitable for dwelling of gophers and narrow-skulled voles.

Table 6. Correlation between the Upper Neopleistocene alluvial and lacustrine sediments of the southern trans-Urals

Age	Horizon	Stage	Alluvium	Lacustrine sediments
Late Neopleistocene	Khanmei (Podporozhskii)	Glacial arid	Bol'shoe Balandino, Bed 6 Miasskoe, Bed 5 Verkhnyaya Alabuga, Bed 5 Streletskoe, Bed 4	Yuzhnyi, Bed 3
		Transitional from glacial humid to glacial arid	Streletskoe, Bed 5	Yuzhnyi, Bed 4
		Glacial humid		Yuzhnyi, Bed 5
	Streletskoe (Mikulinian)	Warm arid	Vvedenskoe, Bed 7	

In addition to species mentioned above, the mammal assemblage from Bed 4 includes remains of *Lagurus lagurus* and *Eolagurus luteus* (Table 3). A high proportion of these remains implies that landscapes included steppe areas with xerophytes, as it is also evident from data of spore–pollen analysis. The steppe lemmings were small, similar in dimensions (Table 4) to those identified in the Streletskoe assemblage (Bed 4) and in the Arapovichi locality of the Russian Plain (Markova, 1982), the latter dated back to the first half of the Late Neopleistocene. According to the enamel coefficient (K = 1.22), the water vole from the indicated bed is close to subspecies *Arvicola terrestris alabugensis* (Maleeva and El'kin, 1986) from the Verkhnyaya Alabuga section. In their morphological features, molars from Bed 4 are similar to those found in the Streletskoe (Bed 4) and Verkhnyaya Alabuga localities.

Small mammals from Bed 3 retain features of the steppe community (Table 3). Their assemblage is remarkable because of absence of *Lemmus* sp. and *Microtus oeconomus*. This suggests more arid conditions that is also evident from the spore–pollen data. The assemblage composition implies that the accumulation period of Bed 3 was characterized by development of steppe and semidesert landscapes in flat interfluve areas with sedge–reed thickets in river floodplains.

Steppe lemmings from Bed 3 are larger as compared with those from the Verkhnyaya Alabuga locality and close in dimension to the specimen from the Miasskoe section (Table 4). Similar dimensions are characteristic of lemming populations of the latest Neopleistocene in the trans-Urals region. In morphological characteristics, molars of steppe lemmings from Bed 3 resemble their counterparts from the Miasskoe locality. The enamel coefficient of water vole molars is close to one (K = 1.08), i.e., it is lower than that characterizing corresponding teeth from Bed 4 and subspecies *Arvicola terrestris alabugensis* (Maleeva and El'kin, 1986). Similar enamel coefficients are typical of Late Pleistocene populations water voles (Agadzhanyan and Erbaeva, 1983).

Table 3 demonstrates the upward (from Bed 5 to 3) increasing proportion of steppe taxa in small mammal assemblages. While rodent remains from Bed 5 lack steppe species, the latter represent 42 and 63% in the assemblages from Beds 4 and 3, respectively. It should be noted also that Beds 5 and 4 yielded fragments of Lemmini teeth as well. In addition, all the described assemblages include species, which can be considered as intrazonal elements.

CONCLUSIONS

Fossils from the Kamyshlov alluvial complex and lacustrine sediments of the Kumlyak Formation enable correlation between the examined sections and elucidate the Late Pleistocene sedimentation history in the southern trans-Urals (Table 6).

(1) The Streletskoe (Mikulinian) interglacial was marked by formation of fluvial facies of the Kamyshlov alluvial complex, which are exposed below or slightly above the water level. Based on fossil remains, the alluvium accumulated during the warm arid interstadial phase (Vvedenskoe locality).

(2) Glacial fluvial facies of the Khanmei (Valdayan) glaciation accumulated in river valleys and in depressions of interfluve areas. The microtheriofauna of this period is lacking steppe species but includes lemming remains. Accordingly, the host alluvium accumulated during the cold humid phase of glaciation.

(3) The cold arid phase of the Khanmei glaciation was responsible for the widespread accumulation of oxbow facies and lacustrine sediments constituting the middle part of the Kumlyak Formation. Corresponding palynological data indicate development of forest-steppe vegetation (meadow–steppe coenoses with pine–birch groves) at that time. Simultaneously, abundance of cryophilic species in the molluscan and ostracode populations increased. Fauna of small mammals implies appearance of steppe and semidesert landscapes and cold arid climate. The latter is also evident from cryogenic deformations and formation of ice

wedges in the uppermost lacustrine clays of the Streletskoe section (Fig. 2). Based on abundant organic remains, oxbow alluvial facies have been considered previously as characterizing the interglacial optimum (Maleeva and Stefanovskii, 1988).

(4) The terminal stage of the Khanmei glaciation resulted in deposition of periglacial alluvial facies and in accumulation of upper lacustrine sediments of the Kumlyak Formation. The structure of alluvial and lacustrine sections suggests that the arid phase was periodically interrupted by cold humid oscillations responsible for appearance of sand and shingle interbeds in the sections. Palynological spectra characterize mainly the steppe landscapes with Chenopodiaceae-*Artemisia* coenoses. The molluscan and ostracode assemblages are of the moderate type with cryophilic elements. The mammal fauna is of the steppe and semidesert appearance.

Thus, comprehensive examination of the Kamyshlov Alluvium and lacustrine Kumlyak Formation of the southern trans-Urals elucidated the detailed characteristics of their reference sections.

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New Data on the Black Sea Level Changes in the Quaternary

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Abstract—Seismostratigraphic study in the outer shelf, an upper part of the Black Sea continental slope, and in the Quaternary depression superimposed on the shelf defined three stages in accumulation of Karangatian sediments, which are divisible into seven episodes. The results are consistent with biostratigraphic subdivision of coastal-marine and continental sequences.

Key words: Black Sea, seismostratigraphy, Karangatian, seismic packet, seismic quantum.

INTRODUCTION

The western Black Sea is well studied by drilling and detailed seismic profiling. Biostratigraphic subdivision of Quaternary shelf sequences is also established. The point of interest is the Karangatian transgression corresponding to the global sea-level rise during the late Neopleistocene interglacial period. Records of this event easily recognizable in the shelf and onland sedimentary sequences are traceable over vast territories of Eurasia. It corresponds to the Riss-Würm in the Alpine scale, the Emsian in West Europe, the Mikulinian interglacial in the European part of Russia, and the Kazantsevo Interglacial in the trans-Urals region (Alekseev and Drushchits, 2001).

The Bulgarian Black Sea shelf and coast with ancient shorelines of this age (Popov and Mishev, 1974) are not an exclusion. According to seismostratigraphic (Limonov and Krystev, 1990; Dimitrov, 2001) and biostratigraphic (Krystev *et al.*, 1990) data, sequences of this age are widespread in the shelf.

It is known that two distinct stages of the Karangatian transgression resulted in formation of two well-preserved terraces 10–12 and 20–25 m high. Altitudes of these terraces are variable, depending on their tectonic settings. The preservation degree of terraces is variable as well. The first late Neopleistocene transgression in the Black Sea was largest and related to the Tyrrhenian transgression in the Mediterranean Sea. According to drilling data, the thickness of Karangatian sediments in the shelf is close to 5 m (Krystev *et al.*, 1990); Fedorov, 2000). At the Caucasian coast, there are three, the Shakheisk, Agoisk, and Sochi terraces corresponding in heights to the Karangatian terraces of the Bulgarian coast. Sediments of these terraces contain diverse molluscan assemblage with species missing now from the Black Sea. Three transgressive phases separated by two regressions of the Karangatian time were distinguished here (Chepalyga, 1997).

The detailed interpretation of seismic profiles in combination with study of coastal deposits and analysis of data on the biostratigraphy and Quaternary stratigraphy of neighboring continental areas provided a new insight into Quaternary sedimentation in the shelf and upper continental slope near the southern coast of Bulgaria.

MATERIAL AND METHODS

Seismoacoustic profiling was carried out in the southern offshore part of the Bulgarian coast of the Black Sea (Fig. 1) during the expedition of R/V *Sredetska-88*. Two areas most informative with respect to sedimentation history were selected for the study. First of them corresponded to the outer shelf and upper part of the continental slope and second one to the Akhtopol' depression, a small sedimentary basin that originated at the beginning of the Quaternary period in the Burgas Basin. The seismic profiling was performed

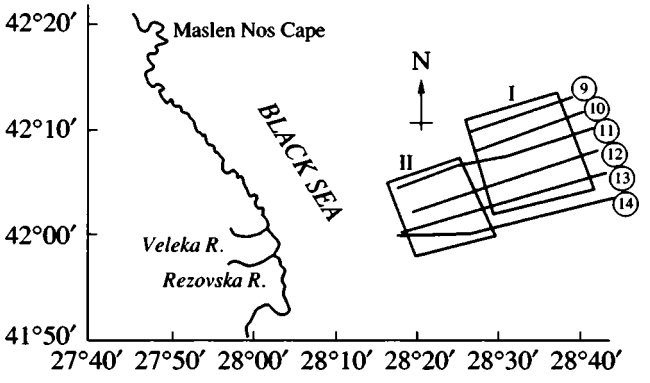

Fig. 1. Scheme of seismic profiles in areas of the outer shelf and upper continental slope (I) and of the Akhtopol' depression (II); encircled numbers denote individual profiles.

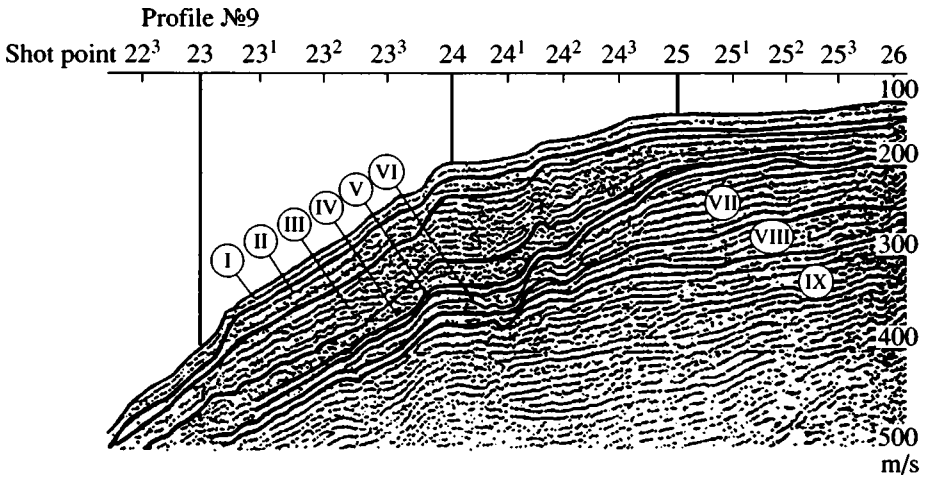

Fig. 2. Seismostratigraphy of Quaternary sediments in the outer shelf and upper continental slope (Dimitrov, 1996): (I, III, V, VII, IX) seismic packets corresponding to transgressive phases; (II, IV, VI, VIII) seismic packets corresponding to regressive phases (see text for age interpretation).

using the seismoacoustic equipment and electric sparker as powerful as of 5 kJ. The total length of seismic profiles is 150 km.

When interpreting seismic records, we used method of seismostratigraphic analysis (Kunin and Kucheruk, 1984). The following units are used in this work: seismic assemblage corresponding to a sedimentary body as thick as dozens to several hundreds of meters or to a period of 0.5 to 5 myr; seismic packet defined using a group of interrelated seismic reflections; and seismic quantum representing a complete single seismic oscillation reflecting both physical parameters of medium and sedimentation event (Kunin, 1989).

The age of seismic complexes and facies was determined using the cyclicity curve that depicts the relative sea-level changes, which are universal and globally synchronous as is postulated in the seismostratigraphic analysis concept. Cycles of sea-level fluctuations can be of local, regional, and global scales (Vail *et al.*, 1982).

SEISMOSTRATIGRAPHY OF QUATERNARY SEDIMENTS

The study area of outer shelf and upper continental slope is characterized by seismic profiles 9 and 10 and by northeastern segments of profiles 11, 12, 13, and 14 (Fig. 1). The Quaternary sequence (seismic assemblage) defined in line with above-mentioned principles is distinguishable in all profiles. Nine seismic packets are established in the sequence (Dimitrov, 1996, 1998, 2001). These are packet IX that can be dated back to the late Chaudan–post-Chaudan, packet VIII corresponding to the post-Chaudan regression, packet VII correlative with the Paleoeuxinian–early Paleouzunlar transgression, packet VI with the time span of post-Uzunlar regression; packet V representing sediments of the Uzunlar transgression, and packet IV fitting to the late post-Uzunlar regression (Fig. 2). The Karangatian layers correspond to seismic packet III. According to seismic profiling, the Karangatian sediments in Quaternary section of the Bulgarian offshore zone are of a high thickness being spread over the shelf and upper continental slope. Sediments of seismic packet II accumulated during the post-Karangatian regression, and seismic packet I represents sediments of the Neoeuxinian and Black Sea transgression.

Seismic packet IV (Fig. 2, Profile 9) formed during the post-Uzunlar regression is of a limited distribution in the shelf, being thicker at the continental slope and having the base that truncates underlying deposits. Reflections are discontinuous, chaotic, and poorly manifested.

In the basal (intervals of 23^1–$23^{2.5}$ and 24^{05}–25^1) and uppermost (intervals of 23^{05}–$23^{2.5}$ and $23^{3.5}$–25^1) parts of seismic packet III corresponding to the Karangatian transgression (Fig. 2, Profile 9), continuous reflectors are brighter and thicker than in other intervals. These features indicate that waves associated with the reflectors proper are of a larger amplitude and period, and corresponding sediments thus differ in physical properties from those in the middle of the packet. Waves with similar parameters are registered also in time sections of other profiles. It seems that seismic records of this kind characterize sediments, which differ in physical properties from other sediments of the sequence. Syn- and post-sedimentation tectonic processes could change properties of sediments with brightest continuous reflectors, while in other intervals they remain unchanged and lacking comparable seismic patterns. This implies that seismic packet III accumulated in the outer shelf and upper continental slope during three stages at least. The first and last stages correspond

probably to long-lasted periods of stable sea level, whereas the middle one characterizes unstable sedimentation conditions. In the slope area, the thickness of the Karangatian sediments with chaotic patterns in seismic records is noticeably lower than in the outer shelf. Closer to the seashore, there are only bright continuous reflections implying that sedimentation developed here under influence of hydrogenic and gravitational factors against the background of subsequent transgressive-regressive cycles.

Seismic packet II corresponding to two regressive phases (Karangatian and post-Karangatian) has a distinct sole unconformable with respect to underlying strata. Changing slope angles of the ancient relief surface and distinctions in properties between basal and roof sediments of the packet are indicative of three different sea-level stands at least during the above regressive phases.

In sediments of the shelf and upper continental slope, the regressive sea-level stands are also well manifested: seismic packets corresponding either to regressive or transgressive cycles have distinct boundaries, and ancient relief forms related to sea level stabilization are clearly seen in seismic records.

The area of the Akhtopol' depression (Fig. 1) is characterized by southwestern parts of profiles 11, 12, 13, and 14. The Akhtopol' structure is easily distinguished in seismic records owing to configuration of reflectors (Fig. 3: Profile 11 (a), Profile 12 (b)). It formed under conditions of regular tectonic subsidence and compensatory sedimentation until the end of the Pleistocene. As a morphological structure, this depression disappeared during the post-Karangatian regression. In the Quaternary period, sedimentary material was transported to the depression from the Western Akhtopol' uplift (Fig. 3a). This uplift represented an island and served as provenance of clastic fraction in sediments of the depression (Limonov and Krystev, 1990). The Western Akhtopol' uplift separates two parts of the study area. The coastal abrasion also produced clastic sedimentary material that was transported by alongshore and contour currents. Thus, the Akhtopol' depression received terrigenous material from the west and east. Interpretation of seismoacoustic records points to the same style of sedimentation during the Quaternary period as well.

During the post-Uzunlar regressive cycle, the flux of sedimentary material from the east was more intense. This tendency continued also at the early stage of the Karangatian transgression (Fig. 3b). To the north (Fig. 3a), influence of the West Akhtopol' uplift on sedimentation in the depression lasted until the post-Karangatian time, as it is evident from distribution of lenticular sedimentary bodies (IV^a, III^a, II^a, III^b, II^b).

Reflectors in profiles 11 and 12 crossing the Akhtopol' depression were numbered for the purpose of precise positioning (Figs. 4a, 4b). The analysis of time records along these profiles (Figs. 3, 4) shows that seismic packet III consists of eleven reflectors (18 to 28), which are not registered in every profile.

The exact distribution parameters of each numbered reflectors is defined as percentage of every reflector length relative to the whole structure length (the Akhtopol' depression) in every seismoacoustic profile. The average parametric values calculated for all time sections across seismic packet III are shown in Fig. 5. As a result, spatial distribution of sediments characterized by reflectors 18 to 28 within the Akhtopol' depression is well established.

The analysis of intensity and configurations of quanta, every of which corresponds to a single reflector, led us to the conclusion that seismic quanta 20, 21, and 24 characterize sediments accumulated during tectonic subsidence episodes, which were not synchronous with the general subsidence of the Akhtopol' depression. These subsidence events involved surfaces significantly smaller in dimensions than the underlying horizon.

It is noted that the surface above seismic quantum 18 represents an unconformity resembling the basal overlap (Mitchum *et al.*, 1982) above seismic quanta 19, 22, and 25. The surface above reflector 25 is a basal overlap relative to seismic quanta 26, 27, and 28 (Fig. 4). Assuming that most and less widespread reflectors correspond to stable and unstable sedimentation conditions, one can distinguish seven episodes of sedimentation within the seismic packet III. They correspond to formation time of (I) reflectors 18, 19, and 20; (II) reflector 21; (III) reflector 22; (IV) reflector 23; (V) reflectors 24 and 25; (VI) reflector 26; and (VII) reflectors 27 and 28.

During the Karangatian time, sedimentation environments in the Akhtopol' depression changed several times. Four episodes of stable sea-level stand subdivide the depression sequence into three stages corresponding to three groups of reflectors 18–22, 22–25, and 25–28. These stages include shorter fluctuations in the sedimentation regime.

At least three episodes of changeable sedimentation can be recognized within the Karangatian and post-Karangatian regressive phases (reflectors 29–35) (Figs. 4a, 4b). Fluctuations in sedimentation environments are also defined for the post-Uzunlar regressive phase (reflectors 12 to 17).

DISCUSSION

The analysis of seismic records along profiles crossing different geological structures of the study region reveals a similarity in changes of sedimentation conditions. Data of seismic profiling in the Akhtopol' sedimentary basin and shelf-break zone show that characteristic of these areas were decelerated rates of the sea-level rise and its stabilization at the initial and terminal stages of the Karangatian transgression. This is consis-

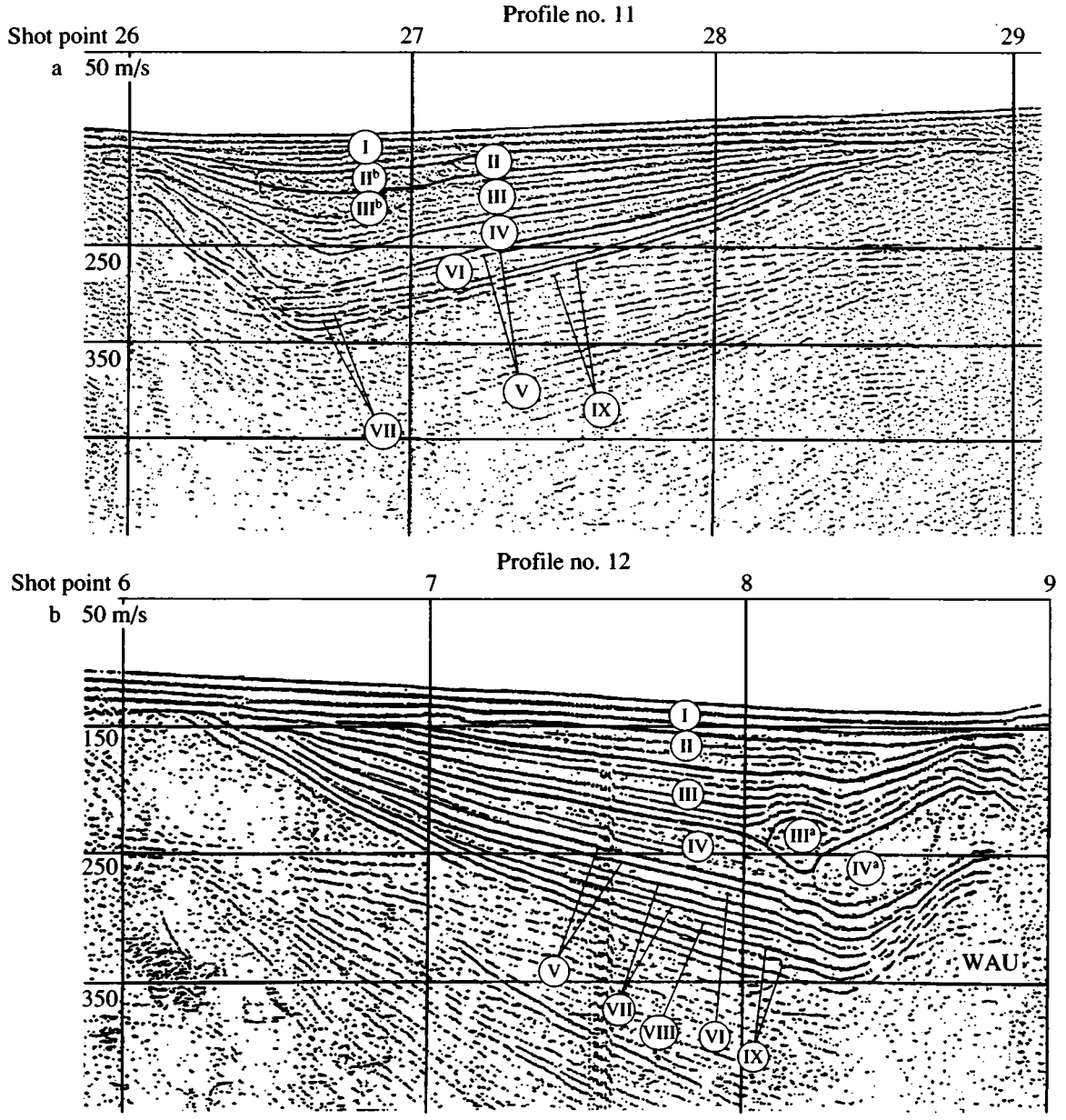

Fig. 3. Seismostratigraphy of Quaternary sediments in the Akhtopol' depression and West Akhtopol' uplift (WAU) after Dimitrov (2001), profiles 11 (a) and 12 (b); encircled numbers denote seismic packets.

tent with the development trend of natural processes that is recorded in many sections characterizing the first late Neopleistocene interglacial period. Sections corresponding to the Mikulinian interglacial in the European part of Russia reflect frequently two warm epochs separated by the colder one. In the lacustrine section of the Protva River valley, sediments of the Mikulinian interglacial are subdivided into six palynological zones. The second half of this time period was marked by development of spruce forests reflecting gradual cooling (Rychagov *et al.*, 1989). The palynological study of the Kelkovo section in the Neva Lowland, where it is composed of marine black clays with the Lusitanian molluscan fauna, revealed two palynological zones in its Mikulinian strata. Pollen from the lower zone represents conifers with an admixture of broad-leaved hornbeam, nut-tree, oak, and elm. The upper zone is characterized by a sharp abundance increase of pollen of broad-leaved trees: hornbeam, nut-tree, oak, and subordinate sycamore, lime, and elm. The data on climate of the Kazantsevo interglacial in East Siberia, eastern Arctic shelf, and the Far East of Russia imply that most complete and well-studied sections of that time reflect three climatic phases: optimum at the beginning, cooling in the middle, and certain warming at the end of the interglacial. The early phase (130–120 ka ago) corre-

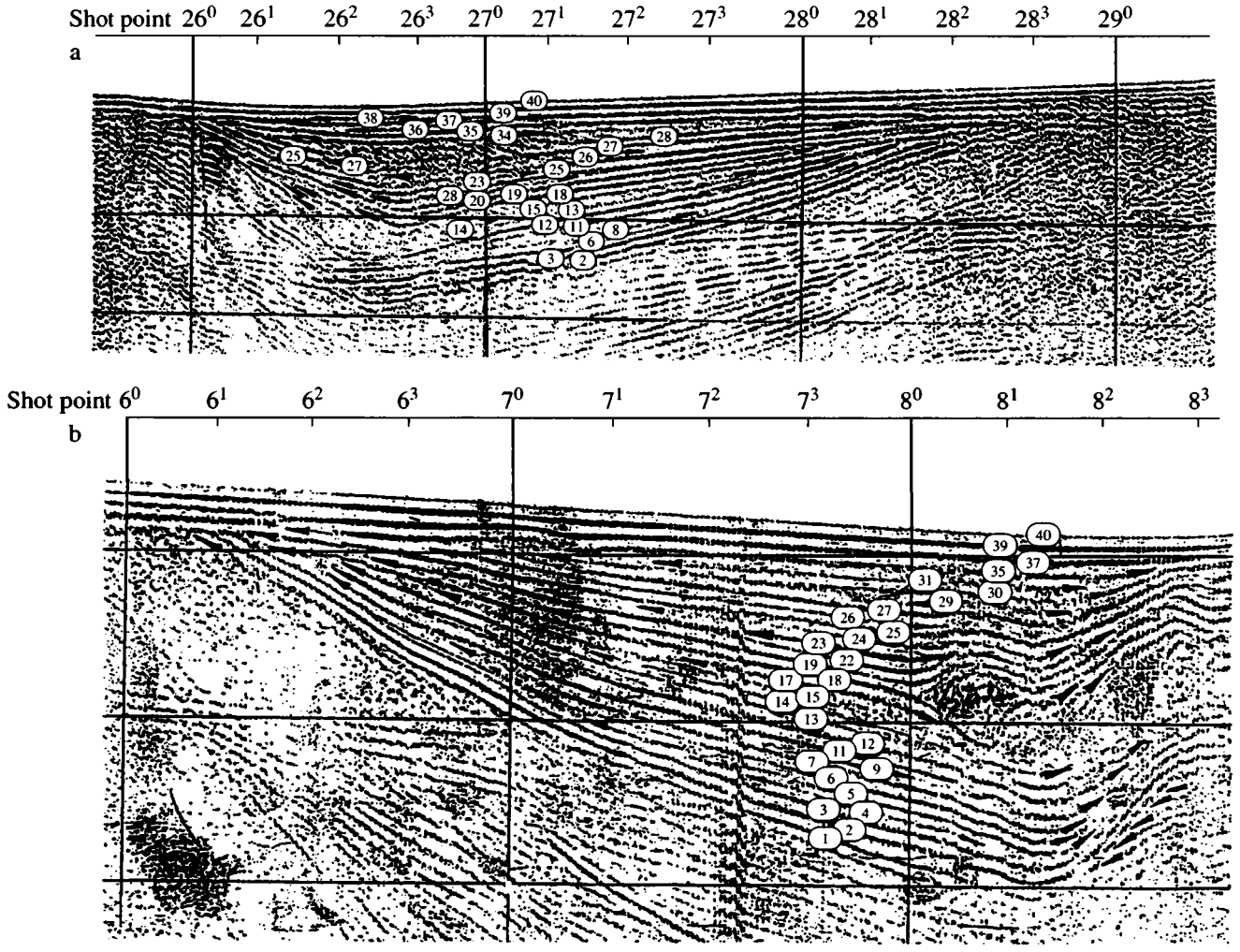

Fig. 4. Reflectors numbered in profiles 11 (a) and 12 (b).

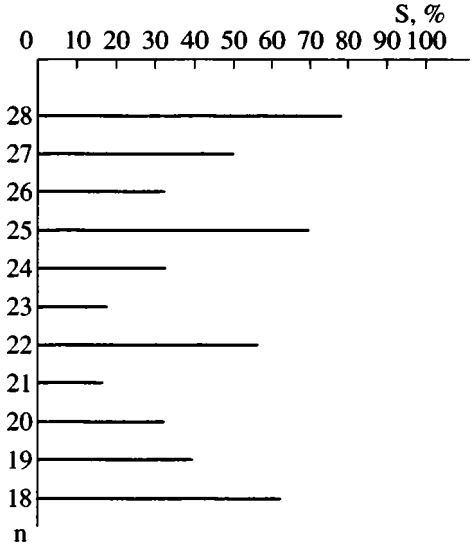

Fig. 5. Percentages of seismic quantum lengths (n) relative to total cross section of the Akhtopol' depression (S).

sponds to Substage 5e of the oxygen isotope scale (Alekseev and Drushchits, 2001). Fedorov (2000) defined three stages in development of the Karangatian transgression: the early and late transgressive phases separated by the short-term regression.

These observations can be correlated with results of geomorphologic studies on land. Two terraces of the Bulgarian Black Sea coast were formed during the Karangatian transgression (Popov and Mishev, 1974). They are situated 20–25 and 8–14 m above the present-day sea level.

Judging from the seismic records, the Akhtopol' depression started to develop at the beginning of the Quaternary and was never subjected to erosion. Even during regressions, this depression probably remained as a local basin or an area of sediment accumulation in the desiccated shelf, and its sequence preserves records of many events of sedimentation history. Seismic quanta defined in seismic records enabled the detailed

stratigraphic subdivision of shelf sediments in southwestern part of the Black Sea.

Based on the results of palynological studies in the Russian Plain, Grichuk (1961) established eight zones during the first interglacial period of the late Neopleistocene. In the coast of Abkhazia, the eastern Black Sea region, seven climatic phases, four warming and three cooling episodes included, were recognized based on palynological data for core samples characterizing the Riss-Würm Interglacial in the depth interval of 38.5–45.0 m (Kvavadze and Rukhadze, 1999). Naturally, it would be incorrect to correlate directly the biostratigraphic data on continental sequences with seismostratigraphic data on shelf sediments. The dependence of shelf sedimentation on the low-amplitude and short-term climatic fluctuations is a problem for further study. It is quite clear, however, that the high-resolution seismic investigations provide an opportunity for detailed stratigraphic subdivision of shelf sequences.

CONCLUSIONS

The seismostratigraphic analysis of shelf sequences is able to define separate small-scale horizons corresponding to the main events in the Pleistocene history of the Black Sea.

Correlation of seismostratigraphic subdivision in shelf deposits of the western Black Sea with the Quaternary stratigraphic schemes established for coastal and surrounding on-land sedimentary sequences allows the following conclusions to be drawn:

(1) Seismostratigraphy based on seismic profiling in the key shelf areas is comparable in resolution with the biostratigraphic subdivision of coeval continental sections;

(2) Notions seismic packet and seismic quantum should be used for characterization of deposits in terms the step-level and more detailed stratigraphic units;

(3) Deposits of the Karangatian transgression that took place in the first interglacial period of the late Neopleistocene are subdivided, like many coeval Eurasian sequences, into three seismostratigraphic units. The units were formed during two warm climatic phases separated by the relatively long-lasted insignificant cooling. This global Mikulinian-Karangatian warming consisted of short-term climatic fluctuations toward warmer and colder climate with different moistening. Seven episodes of this kind can be defined in the Karangatian sequence of the southwestern Black Sea.

Reviewer M.N. Alekseev

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In Remembrance of Michael A. Pevzner (1937–2003)

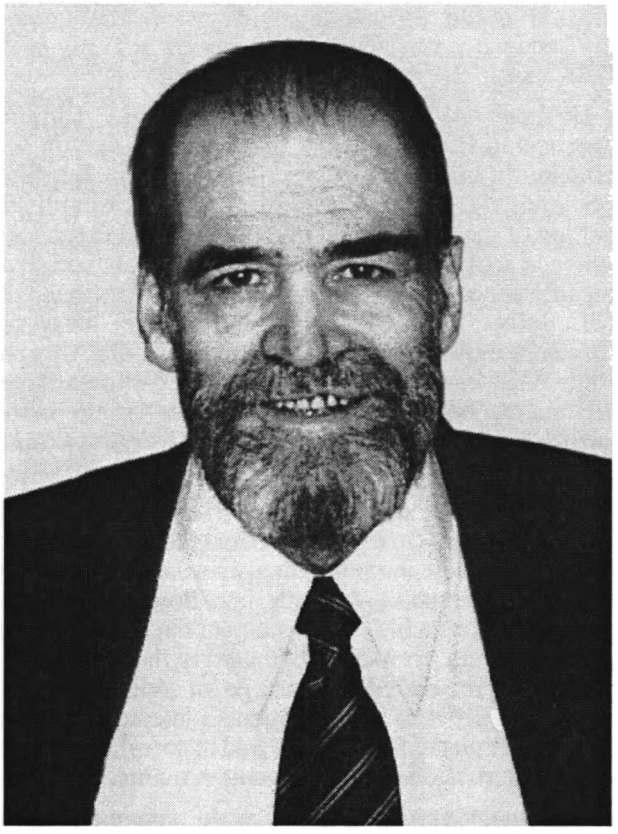
We lost Michael A. Pevzner, our good friend and colleague, a nice personality and talent scientist. He was one of first soviet experts in paleomagnetism, an inexhaustible toiler, disciple and workmate of V.V. Menner and K.V. Nikiforova, the establisher of Paleomagnetic Laboratory at the Geological Institute, Russian Academy of Sciences, and initiator of investigations in that field.

M.A. Pevzner was born April 2, 1937, in Leningrad, where his parents worked for the All-Union Research Institute of Geology. Abram S. Pevzner, his father who graduated from the Minsk University was among initiators of emanation survey of uranium ores and worked in the Transbaikal region. After the World War II he was teaching mathematics and astronomy. Elizaveta L. Abakumova, his mother, worked as mapping geologist in the southern Urals and contributed much to development of iron-and-steel industry in the Magnitogorsk area. During the wartime, the family was evacuated to the Altai, where life was hard. In his memory, it was a time of everlasting hunger, when even a salt grain seemed to be dainty. Being a child, Michael must take care of his younger sister when parents went to work and older brother and sister were at school. It was a source of diligence, sociability, and kind attention to people, which were characteristic of him.

M. Pevzner became interested in geoscience when he was a schoolboy. He spent his summer holidays in geological, archeological, and botanical expeditions to the Pazyryk Mounds, Altai, and Pamirs. He was fond of excavations, sampling, and mineral collecting in the field. In 1955, he left school in the Gorno-Altaisk to become a student at the Leningrad Mining Institute and graduated from the latter as mining engineer-geophysicits in 1961. He got first job in the town of Tura (Krasnoyarsk krai) where worked for the Expedition 20 organized by the 6th Department of Minisry of Geology and Interior Protection of the USSR.

Scientific career of M. Pevzner begun in 1962 at the Geological Institute, USSR Academy of Sciences, where he rose from the rank of technician to the leading scientist. Geography of field trips by M. Pevzner is extremely broad: Kamchatka, Buryat Republic, Tajikistan, Azerbaijan, Georgia, Northern Caucasus, Taman Peninsula, Crimea, Southern Ukraine, Moldova, Trans-Carpathian region, Belarus, Lithuania, Austria, Hungary, Iceland, and India.

Kamchatka was first large region for paleomagnetic study by Pevzner. He originated the idea and elaborated method of paleomagnetic mapping first tested in 1968



when the VAGT Expedition 9 carried out geological survey of Kamchatka at the scale of 1 : 200000. Paleomagnetic measurements of hundreds oriented rock samples were used to trace the 0.7-Ma-old boundary over the area as large as 13000 km^2. As consequence, the regional stratigraphy of volcanogenic rocks and their accumulation rates were elucidated. Later on (1971–1973), the Ministry of Geology included the Pevzner's method of paleomagnetic mapping into the Refresher Course Program at the Moscow State University and recommended to use it in the geological survey areas with widespread Cenozoic volcanics.

M.A. Pevzner got his first academic degree in 1968 after the candidate (geol.-min.) dissertation "Paleomagnetism and Some Problems of Pliocene-Quaternary Stratigraphy in Kamchatka." His paleomagnetic study in areas around the Sea of Azov contributed much to the Quaternary chronostratigraphy in the Eastern Europe. At that time, there were initial attempts to correlate geological events in the glacial, periglacial, and maritime zones of the East European plain based on the

reference loess–soil succession. In the period of 1968 to 1970, Pevzner and his colleagues studied a series of sections at different geomorphological levels in the Sea of Azov region. Afterward, using the detailed paleomagnetic and facies analyses, he established position of the Brunhes–Matuyama reversal in the system of loess–soil sequences and outlined stratigraphic relations between the continental and marine deposits. This magnetostratigraphic investigation of complicated Quaternary sequences was actually pioneering not only in the USSR, but in the rest of Europe as well.

He continued his research in Moldova, Ukraine, Austria, and Hungary. Work in collaboration with academician Márton Pécsi from the Hungarian Academy of Sciences lasted from 1972 to 1991. They established position of the Brunhes–Matuyama reversal in the loess sequence and elucidated paleomagnetic characterization of Pannonian deposits in Hungary. Together with Prof. Julius Fink, Pevzner studied the loess–soil sequences in the northeastern foothills of the Alps and in the Wien basin of Austria. In 1976, he investigated paleomagnetism of the Upper Sivalik Group in the North Himalayas and the Narmada River alluvium in northern and central India.

In 1978-1979, M. Pevzner worked in Iceland. He was first to study here the paleomagnetic structure of 150-m-thick volcanogenic–lacustrine deposits below the thick sequence of volcanic lava flows. His results, when correlated with paleobotanical data, showed that lacustrine strata, formerly attributed to the Pleistocene, are in fact of the early Pliocene age. In addition, he used paleomagnetic data on Neogene plateau basalts to define age limits of Subchron 9 and to correlate sections of volcanic rocks barren of organic remains.

The magnetostratigraphic scale spanning range from the Sakaraulian to Apsheronian regional stages of the Black Sea and Caspian regions was elaborated by Pevzner during the initial 1980s after his investigations in Azerbaijan, Georgia, Northern Caucasus, Taman and Kerch peninsulas. Together with V.N. Semenenko, he defined limits of the Pontian regional stage in the magnetochronological scale to show the short-term range of this subdivision important for stratigraphy of the Eastern Paratethys and thus specified stratigraphic ranges of conjoined regional stages. Under focus of his attention, there were also middle Miocene to Pleistocene deposits of southern Europe, Mediterranean region, Eastern and Central Paratethys, especially their correlation with the magnetochronological scale and with biozonations established in oceanic sediments. In 1986, Pevzner got the next academic degree for the doctorate dissertation "Middle Miocene to Pliocene Stratigraphy of Southern Europe Based on Paleomagnetic Data," in which all his results have been summarized.

These results were of a high scientific importance even for regions, where he did not work personally. For instance, they have been used in the Cenozoic magnetostratigraphic scale elaborated for Mongolia based on combination of geochronological (K-Ar, TL) and biostratigraphic data. This scale represented a solid stratigraphic framework not only for regional correlations in Inner Asia, but also for global geological reconstructions.

Until the last moments of his life, M. Pevzner was enthusiast of magnetochronological study of Neogene to Quaternary burials of mammal remains. This approach was used to verify geological time spans of important stages in evolution of mammals and to enhance resolution of mammal zonations established in the continental Neogene of Europe. His contribution to paleomagnetic characterization of mammal burials in Moldova, Ukraine, Georgia, Azerbaijan, Northern Caucasus, Sea of Azov region, Central Asia, West Siberia, and Hungary is extremely valuable. Applying methods independent of paleontological principles, he correlated reference localities of mammal bones in southern Europe, Siberia, Central Asia, and Transbaikal region. This work considerably corrected former viewpoints and elucidated many aspects of paleozoogeography and fauna evolution in different provinces and landscape–climatic zones.

A fitting testimony to his scientific achievements in recent years is biometric method of dating the mammal remains. Pevzner elaborated this method based on the established mathematical equation describing the time dependence of evolutionary significant morphological parameters of mammal remains, such as hypsodonty of rhizodont voles, dental plate frequency in the mammutid lineage, cranium sphericity coefficient of fossil man, etc. He was also involved into the project of elaborating the detailed stratigraphic scheme for the continental Neogene–Quaternary deposits in collaboration with experts in paleontology. As a part of the scheme, the Middle Pliocene–Quaternary mammalian zonation specified better than others was calibrated with reference to the magnetochronological scale, and the established zones turned out to be well traceable from England to the Transbaikal region.

It was very typical of him to analyze comprehensively any problem he met. Using radiometric dates, he found it necessary to get a deep insight into all methods of age determination. With respect to the continental stratigraphy, he deemed it wise to know everything important in morphology of mammal remains and in evolution of every faunal group. Referring to earlier geological or paleomagnetic concepts, he never have mentioned them simply, but always with a critical analysis of their validity. It was easy to work with him even having a controversial viewpoint on the issue, because he was open to reasonable arguments and tolerant to criticism.

It was something special the attitude of Pevzner to younger workmates. He was always ready to give a good advice, to read a manuscript or a dissertation, being attentive to any work of whatever kind it would be, on geology, palynology, foraminiferal or mamma-

lian stratigraphy. His comments were meticulous, constructive, objective, and always benevolent. He was a persistent member of the Bureau on Neogene and Quaternary Stratigraphy of the Interdepartmental Stratigraphic Commission, a well-known expert on the Neogene stratigraphic schemes submitted to the commission. In this work, he was also meticulous, patient, and unbiased.

M. Pevzner was gifted by nature with a special organizing ability, as one can judge from his activity in connection with the international scientific meetings, e.g., with the INQUA Congress of 1982 or with the International Geological Congress of 1984 held in Moscow. He was able to plan the field work so neatly that it was comfortable, as far as it possible, to investigate many objects in a quiet friendly atmosphere, without unnecessary difficulties. Arriving to a new place, he quickly became banded by sympathy with inhabitants, and many of them turned into his friends forever. He was able to find a common language with anybody he met, from an academician to a simple farmer. The social status of a good man was not important in his opinion.

M. Pevzner was a man of unusual heartedness, kind and generous, with a keen susceptibility of injustice and aggressive misunderstanding. His first reaction was always to help people when they need in support. Many colleagues and friends saw how he rushed into a dangerous situation in order to right a wrong whatever it would be a scientific or social event.

It seemed that he has enough energy for everything, even for the arduous tusk of interdisciplinary investigations in collaboration with geologists, anthropologists, biophysicists, and botanists. He was an active participant of anthropological seminars organized by Russian Academy of Sciences, acted as opponent to doctorate dissertations on various problems of anthropology.

Michael A. Pevzner married Galina M. Proskuryakova, a nice and wonderful woman. They lived in a full harmony, taking care of each other all the life. It was always warm, homelike, and jovially in their hospitable household. Michael was a perfect family man very attentive to his wife, daughters, and grandchildren. A scientist well known in the world he was unpretentious nice man in the common life, housewifely and deft, with hands of gold.

He was the only one among us gifted so generously with the warming soul, vital energy, lambent humor, indefatigable diligence, and great talent. His untimely death is a very sad shock to all who knew him, the grievous loss.

Friends and workmates

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